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COLLAPSE FEATURES, TEMPLE MOUNTAIN URANIUM AREA, UTAH

By PAUL F. KERR, MARC W. BODINE, JR., DANA R. KELLEY, AND W. SCOTT KEYS

ABSTRACT

Hydrothermal solutions mainly of hypogene origin are believed to be responsible for the formation of collapse features, their associated alteration effects, and the accompanying mineralization at and near Temple Mountain, Utah. The collapse features are characterized by: (1) a central core, consisting of blocks and fragments of downdropped broken strata; (2) a structural basin in surrounding strata; (3) marked alteration effects which differ in core and border. The collapse features, 11 of which are noted, range from less than 100 feet in diameter to more than 1000 feet and with three exceptions consist of a Moss Back core in Moenkopi country rock. Even the three exceptions represent a similar process. The writers believe that the collapse structures were formed by the removal of the underlying carbonate rocks (Kaibab Limestone and lower and middle Moenkopi Formation), which resulted in downdropping of the higher strata and sagging of the surrounding beds to form the enclosing basin.

At the Temple Mountain collapse Wingate and Moss Back sandstone blocks have been dropped several hundreds of feet and are enclosed in structural basins of Moenkopi and Chinle strata. Drilling records reveal that the carbonate constituents of the underlying Kaibab Limestone have been removed. Alteration effects within the collapsed core and the surrounding country rock from the top of the Coconino Sandstone through the Wingate Sandstone include argillization, carbonatization (dolomitization), and ferrugination (sideritization and hematitization); the uranium mineralization following consists chiefly of emplacement of pyrite, urano-organic ore, and late alteration effects, chiefly alunitization.

Mineral relationships indicate that hydrothermal solutions dominated in processes of the development of the collapse features and in the alteration and mineralization.

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Introduction

Isolated erosional remnants of Wingate Sandstone rise above the surrounding slopes to form the cathedral-like peaks of Temple Mountain (Pl. 1, fig. 1). The sandstone caps make a distorted H: North Temple Mountain and South Temple Mountain form two sides, and the connecting ridge (Pl. 1, fig. 2) is a broken bar.

The mountain lies along the southeastern side of the San Rafael Swell (Fig. 1) 44 miles in a direct line southwest of Green River in Emery County, Utah. The swell is one of the great structural features of the West and forms a domal uplift which extends southward between Green River and Price, Utah. A steep elliptical rampart of Wingate Sandstone cliffs with a gradual outer slope encircles an elevated undulating surface. The main asymmetric anticlinal fold trends northeastward for about 45 miles.

Southwest of the swell are the Waterpocket fold and the Circle Cliffs which are two similar, smaller asymmetric anticlines. North of Green River, the Roan Cliffs trend eastward, and east of the swell is the Green River Desert, a region of flat-lying sediments drained by the Green and Colorado rivers.

Uranium mining on Temple Mountain (Fig. 2) began with the search for sources of radium in the early 1920's (Hess, 1922; 1923, p. 272–276). Deposits containing uranium and vanadium were worked for a time, but operations were discontinued with the advent of foreign uranium sources. In the search for uranium for the production of fissionable material, the mines were reopened and for several years (Gruner, 1951) have been important sources of uranium.

Temple Mountain appears to be unique among the uranium occurrences of the Colorado Plateau. Here strata in various horizons (Fig. 5) carry isolated bodies of uranium mineralization, but ores are chiefly concentrated in the Moss Back Member of the Chinle Formation. Stratigraphically, the vertical spread between the uppermost (Wingate) and lowermost (Kaibab) ore occurrences is about 1300 feet.

The east and west slopes of Temple Mountain exhibit great blocks of slumped or collapsed strata which lie partly buried. Disturbed zones are in each case encircled by older warped beds that are essentially unbroken. In addition to this occurrence 11 smaller isolated collapse features are known in the southern part of the swell.

The rocks of collapsed areas are bleached

light gray or almost white in contrast to the reds, browns, and tans of surrounding Moenkopi or Chinle strata. The features range in diameter from 100 feet to more than 1000 feet

main assignment of this project. It involved the features and surrounding areas of altered rock, which are significant in connection with the emplacement of the associated uranium de-

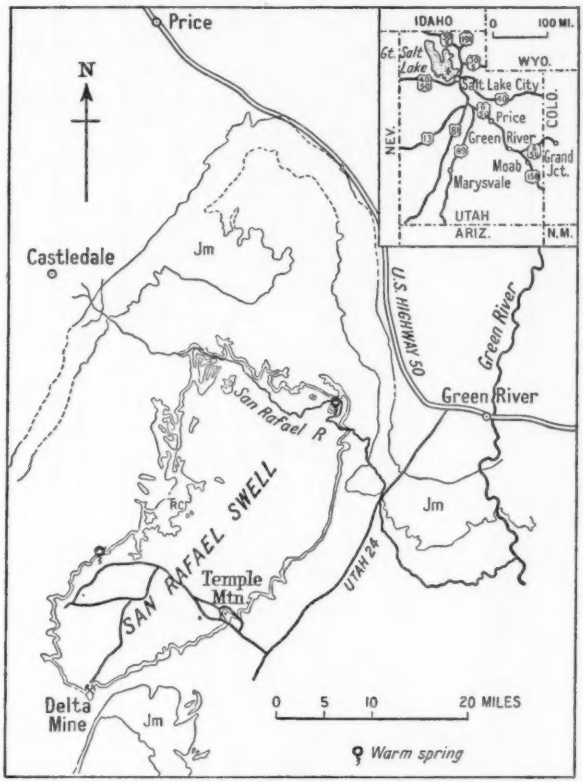

FIGURE 1.—LOCATION DIAGRAM OF SAN RAFAEL SWELL
Outcrop areas of Chinle (Trc) and Morrison (Jm) outline the structure.

and include zones of partial structural confusion.

The most frequently visited collapsed area, referred to by miners as the flopover, lies on the west slope of Temple Mountain (Pl. 1, fig. 1. This feature has been observed by many geologists and has been variously interpreted as a landslide, a breccia pipe, a filled sink hole, and a choked collapse. The senior author, accompanied by Y. William Isachsen, T. W. Mitcham, and Hiram B. Wood, noted this area in the summer of 1952 during visits to the San Rafael Swell.

The study of collapsed features has been the posits. The ores of the mining area are to be covered in another study.

An important aspect of the Temple Mountain problem concerns the pressure-temperature conditions of uranium emplacement. Some authors believe that the uranium originated under the influence of normal ground water. Others have proposed a hydrothermal origin and a syngenetic origin.

The current investigation includes studies by the Sinbad Field Office of the Atomic Energy Commission and follows experience in the observation of hydrothermal uranium-bearing veins at Marysvale, Utah (Kerr *et al.*, 1957).

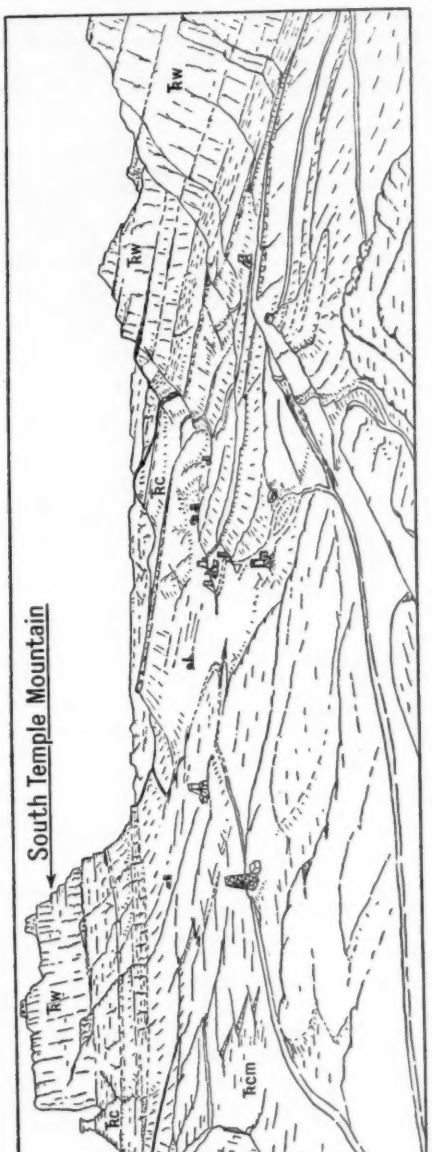

FIGURE 2.—CALYX AREA AND TEMPLE MOUNTAIN
Large Calyx holes (30 inches in diameter) about 100 feet deep serve as mine shafts. Urano-organic ore is mined from Moss Back strata.

References to the Temple Mountain work have been made in a general discussion of the natural occurrence of uranium (Kerr, 1955, p. 5–59) and in connection with alteration criteria (Kerr, 1956, p. 633–639). Field studies covered the summer of 1954 and part of 1955 and were coupled with laboratory work at Columbia University during the winter months of each year. This report is a prelude to work in preparation, which concerns the mechanism of emplacement of the urano-organic ores and the hydrothermal alteration associated with the collapse areas.

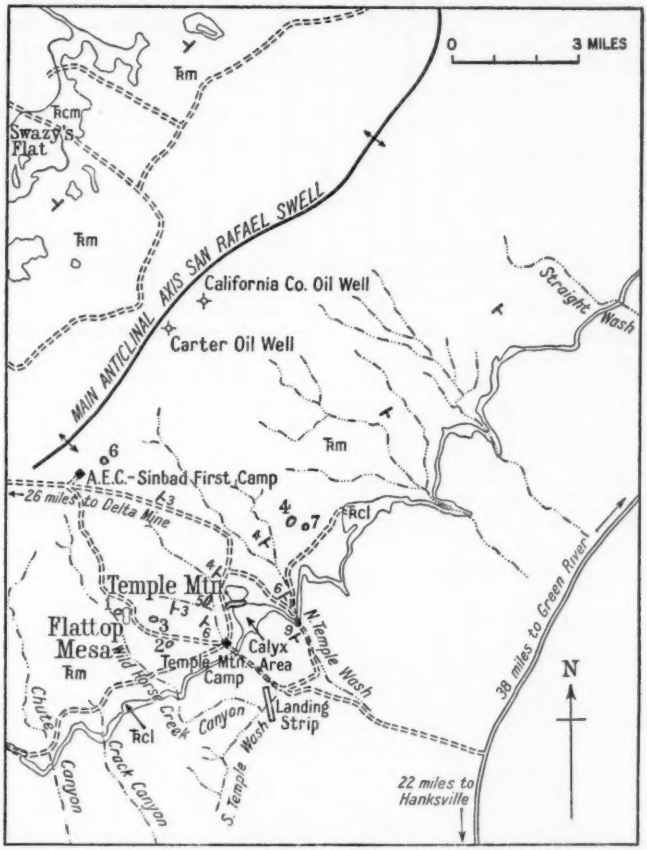

FIGURE 3.—SKETCH MAP OF SAN RAFAEL SWELL NEAR TEMPLE MOUNTAIN
The formations shown are Moenkopi (Trm), Moss Back (Trcm), and undifferentiated Chinle (Trcl).

Structural Setting

Temple Mountain lies about 6 miles southeast of the main anticlinal axis of the San Rafael Swell (Fig. 3). The crest is so broad that in places the precise axis is poorly defined, but the prevailing trend northwest of Temple Mountain is about N.45°E. About 12 miles to the north it changes to more nearly north.

The northeastern flank of the swell is asymmetric and has steep dips, some of which are as high as 85°E. Lower dips prevail elsewhere. Near Temple Mountain the general southeast attitude ranges from 4° at the north end of the mountain to 7° on the south edge of the mining area.

Faults near Temple Mountain (Fig. 4) strike about east and are nearly vertical. The throw ranges from several feet to several tens of feet and shows considerable range along the strike. Only a few faults are more than 1 or 2 miles

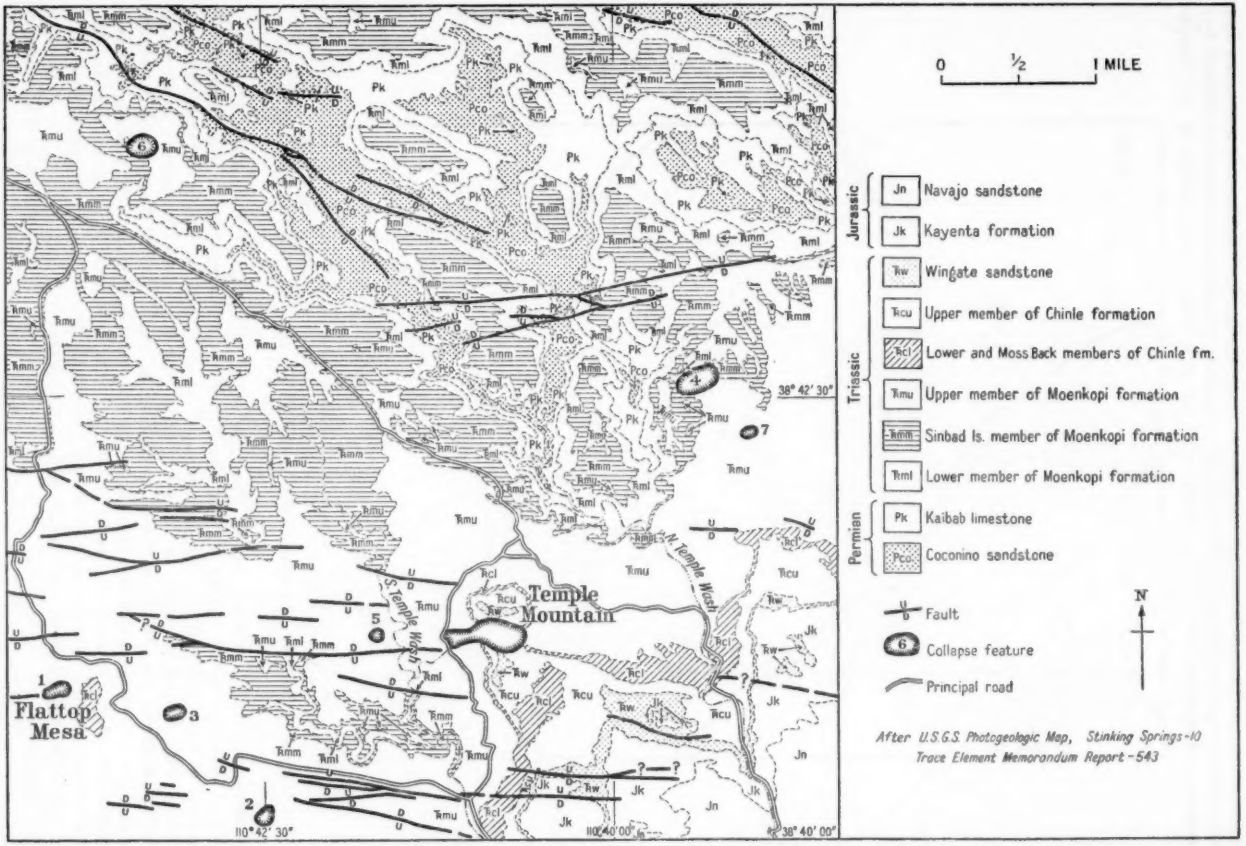

FIGURE 4.—GEOLOGIC FEATURES AND COLLAPSED AREAS NEAR TEMPLE MOUNTAIN

long. Certain east-west joint systems may indicate concealed underlying faults in strata below.

Two strong sets of joints appear on rocky dip slopes of Wingate or Navajo sandstone on the outer slope of the uplift. The stronger set is parallel to the strike of the bedding, and the weaker set is nearly at right angles. Both sets are apparently related to the uplift of the anticline. Urano-organic ore forms lenses near fractures and faults southeast of Temple Mountain.

Igneous Features

Igneous features of the San Rafael Swell are not well known. Dikes, sills, minor stocks, and small volcanic craters occur and are well shown on air photos of the southwestern outer slope of the swell. Dikes also invade the south end of the swell. The more prominent adjacent bodies have been mapped by Gilluly (1929, p. 69–130), who has also furnished a petrographic description (1927). These features are also shown on the photogeology sheets of the U. S. Geological Survey.

The California Company (Hatch, 1941) drilled to Precambrian gneiss, which was encountered at a depth of 4360 feet. The well is located in Emery County, Utah, T.23 S., R.11 E., sec. 27, in the NW ¼ of the NW ¼, along the axis of the swell about 7 miles N. 10°W. from Temple Mountain.

The core at 4400 feet (Hatch, 1941) is foliated. It has a schistose appearance and contains chiefly plagioclase (An 48), orthoclase, microcline, quartz, biotite, and garnet and less abundant pyrite, rutile, apatite, and zircon. The order of crystallization of the gneiss appears to be typical of a granodiorite, and the texture confirms an igneous origin.

The Henry Mountains (Hunt, 1953) lie south of the San Rafael Swell and are underlain by diorite porphyry stocks which are surrounded by laccoliths of similar composition. These igneous rocks occupy the center of a large synclinal basin.

Near by, to the south and west, are the lava-covered slopes of the High Plateaus of Utah, which enclose the Marysvale uranium district. In this district volcanism has been accompanied by igneous intrusion and widespread hydrothermal alteration, and clay bodies associated with uranium have developed (Kerr *et al.*, 1957).

Notwithstanding these features, no direct connection between igneous activity and the uranium ores of Temple Mountain has been established.

Acknowledgments

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The field studies benefited from suggestions by Dr. Walter H. Bucher, Columbia University. The writers were assisted during part of the summer by Mr. Hans W. Schreiber, also of the Department of Geology at Columbia. Courtesies extended by the various mine owners and operators, particularly Mr. W. Moulton, superintendent of the Consolidated Uranium Temple Mountain mines, are sincerely appreciated. Miss P. K. Hamilton, Department of Geology, Columbia University, has aided considerably in the laboratory studies and has given many suggestions concerning mineralogical problems.

Stratigraphic Relations

Pre-Permian Strata

Information concerning the strata below the Coconino (Fig. 5) is limited. A well drilled by the California Company along the anticlinal axis penetrated Precambrian gneiss. The surface formation at the drill site is the Sinbad Limestone Member of the Moenkopi Formation about 100 feet above the base. Where the well was drilled the strata are about horizontal; hence about 4000 feet of Permian, Pennsylvanian, Mississippian, or older strata were penetrated.

Stratigraphic data on formations beneath the swell are incomplete. Wengerd and Strickland (1954), in sections taken near by, show the Hermosa, Paradox, and Leadville formations beneath the Cutler or Coconino. The Hermosa Formation (Pennsylvanian) below the "R" datum would be about 1200 feet thick, the Paradox, Pinkerton Trail, and Molas formations (Pennsylvanian) about 600 feet thick, and the Leadville Formation would have an unknown thickness. These thicknesses plus the Coconino and Kaibab as reported by Gilluly (1929) total less than the sedimentary strata reported in the California Company well.

The driller's log of the Carter Oil Company

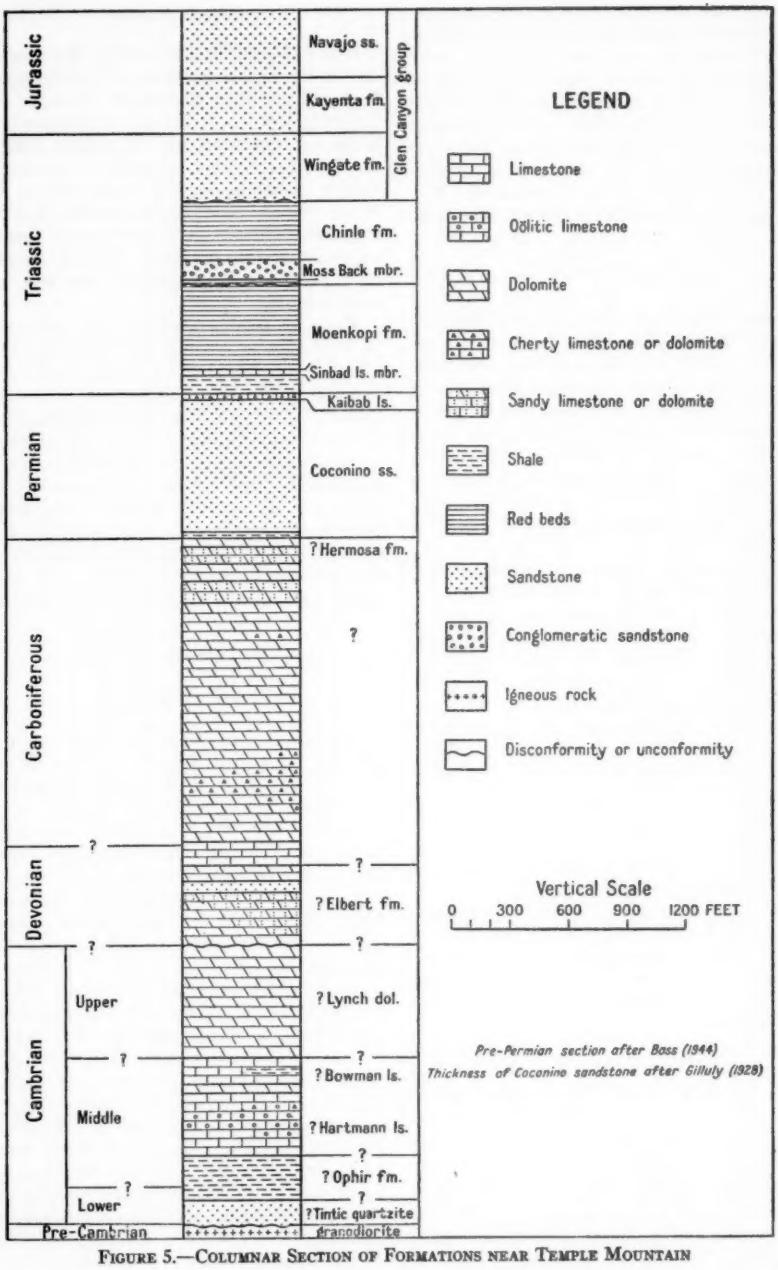

FIGURE 5.—COLUMNAR SECTION OF FORMATIONS NEAR TEMPLE MOUNTAIN

well in T.23 S., R.11 E., sec. 34, to a depth of 3035 feet is given by Gilluly (1929). This location is about 5–5½ miles almost due north of Temple Mountain (Fig. 3). Presumably the lower 2000 feet of the well penetrated Pennsylvanian beds and apparently began near the top of the Permian (Baker, 1935).

Permian Strata

Coconino Sandstone.—The Coconino Sandstone of the swell has been correlated with the Coconino of the Grand Canyon region (Gilluly, 1929) on the basis of lithology and stratigraphic position beneath marine Permian limestone. To the east strata of corresponding age have been referred to the Cutler Formation (Baker, 1935; 1946; Wengerd and Strickland, 1954). The Coconino Sandstone is exposed for a thickness of 713 feet in Black Box Canyon of the San Rafael River (Gilluly, 1929). Near Temple Mountain only 50 feet has been exposed. An unusual feature of the Coconino here is the occurrence of dead-oil sand. Similar sand in greater thickness occurs along Straight Wash, about 4 miles northeast. Petroliferous Coconino sand was also found in a churn drill hole in collapse no. 4, a short distance east. Coconino in diamond-drill holes on Temple Mountain was penetrated for 50 feet. It was also petroliferous. In most places the Coconino Sandstone is tangentially cross-bedded, gray, fine- to medium-grained, and cemented with calcite. It is probably aeolian in origin.

Kaibab Limestone.—Marine Kaibab Limestone, about 30 feet thick, is exposed about 1 mile north of Temple Mountain. It ranges in thickness from 0 to 90 feet 13 miles southwest of Temple Mountain. The contact between the Kaibab Limestone and the underlying Coconino Sandstone is gradational. The base of the Kaibab consists of limy siltstone and sandstone which partly represents reworked Coconino. Most of the formation is a dark-gray to black limestone which is commonly sandy. Chert concretions with calcite and asphalt cores are abundant in the upper Kaibab. North of Temple Mountain the limestone appears to contain more asphalt than is usual. It occurs in vugs and between the grains.

An abrupt change from full thickness to absence along strike may be attributed in places to thermal leaching near faults. The extent of thermal-solution effects in contrast to old erosional features and gaps in sedimentation warrants further study.

Triassic Strata

Moenkopi Formation.—A noticeable erosional unconformity separates the Kaibab Limestone from the overlying Lower Triassic Moenkopi Formation. A basal conglomerate, about 5 feet thick, contains many chert and limestone fragments which are derived from the Kaibab Limestone. North of Temple Mountain the Moenkopi Formation is only 556 feet thick, which is less than other thicknesses in the swell (Gilluly, 1929).

The Moenkopi Formation consists mainly of red to brown shaly siltstone and fine-grained sandstone in lenticular beds that range from 0 to 10 feet thick. The sandstone is commonly cross-bedded and contains abundant ripple marks. The Sinbad Limestone Member consists of 30 feet of buff to gray thin-bedded limestone strata, approximately 90 feet above the base. It is of marine origin, whereas much of the rest of the Moenkopi was probably deposited in shallow lagoonal waters amid arid surroundings.

The color of the Moenkopi shows marked lateral variation, and sharp color boundaries transgress bedding planes in many places. Large gray and buff areas may have resulted from the alteration of normal red sediments. Color changes along strike are in places abrupt and appear related to structural features rather than facies changes. Distinctly local altered zones are recognized along certain faults.

Chinle Formation.—Considered Late Triassic (Gregory, 1917), the Chinle has been extensively studied. The terminology represents local field usage and may be modified as more extended studies become available. The lowermost Chinle (Fig. 6) which heretofore has been assigned to the Moenkopi consists of 25–40 feet of gray and maroon, varicolored siltstone. This unit locally consists of three parts and is subject to pinto alteration. It shows no angular discordance with the underlying Moenkopi, although it lies above a disconformity. It is included in the Chinle because it lies above the disconformity and is lithologically similar to a greater thickness of similar Chinle strata above the Moss Back Member.

Robeck (1956) recognized the continuity of this unit throughout the southern half of the swell. He applied two member names in the Temple Mountain area. The upper part (the two units below the Moss Back in Figure 6) is

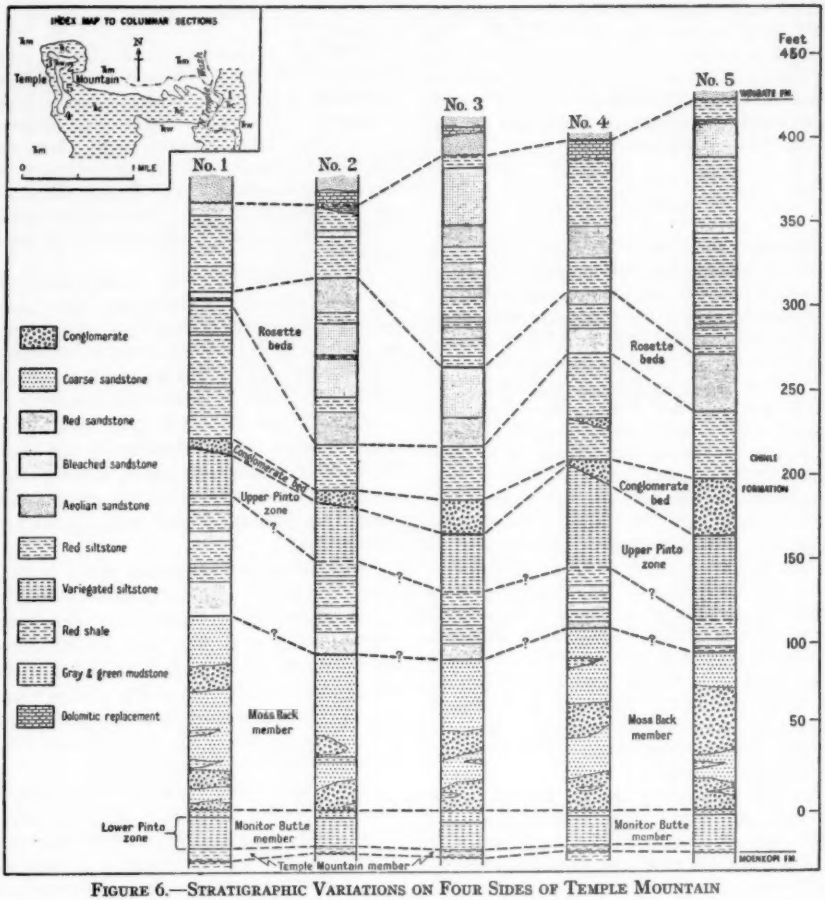

FIGURE 6.—STRATIGRAPHIC VARIATIONS ON FOUR SIDES OF TEMPLE MOUNTAIN

the Monitor Butte Member, and the lower part just above the Moenkopi is the Temple Mountain member.

The Monitor Butte forms the most uniform and continuous stratum in the vicinity of Temple Mountain (Fig. 6). It is separated from the Moss Back above by a thinly laminated, fine splintery shale. The latter weathers to purple red in contrast to the brick red of the Moenkopi below.

The pinto zone appears to range from a mottled white gray brown to almost white. It is white near the Vanadium King workings, where bleaching is localized along a fault zone. It is a significant location, since it lies below productive portions of the Moss Back which have been mined.

In most places the Monitor Butte is unbleached, but bleaching occurs close to the collapse on Temple Mountain and in places along faults.

This feature may represent the effect of hydrothermal solutions.

The thick ore-bearing sandstone previously has been called the Shinarump Conglomerate, but stratigraphic work by geologists of the U. S. Geological Survey (Robeck, 1956) suggests that it is correlative with the Moss Back Sandstone of White Canyon. The Moss Back is 50–125 feet thick at Temple Mountain and averages 100 feet. The lithology is variable, but the largest part of the unit consists of limy, poorly sorted, gray to brown sandstone and some conglomerate. The conglomerate is made up

primarily of lime and chert pebbles. Some plant remains occur in the sandstones in the form of logs. Cycad structure has been identified in some Chinle plant remains from the swell. Abundant gray-green mudstone lenses are intercalated throughout the Moss Back. Festoon or trough cross-stratification is common.

The Chinle above the Moss Back ranges from 185 to 265 feet thick and averages 200 feet. It is lenticular, and in this vicinity few units are traceable for as much as 2 miles. Rock types in the upper Chinle in decreasing order of abundance are: siltstone, sandstone, mudstone, and conglomerate. Aside from the conglomerate, which is gray, these rocks are normally red, brown, and variegated purple. Bordering the collapse areas at Temple Mountain they are locally light gray, brown, and greenish because of alteration. A mottled effect similar to the pinto rock below the Moss Back occurs in places. Plant remains are found in the coarser units, and some uranium-vanadium ore occurs near the main collapse. Most of the Chinle Formation of the swell is thought by the writers to be of fluviatile origin.

Glen Canyon Group

Wingate Sandstone.—The Committee on Geologic Names (February 1954) assigned the Wingate Sandstone to the Triassic System, but the rest of the Glen Canyon Group is still considered to be Jurassic. Recently Averitt *et al.* (1955) showed that the Wingate Sandstone of the Kanab, Utah, section is not correlative with the Wingate to the east and renamed it the Moanave Formation. Whether this applies to the San Rafael Swell has not been determined. The Wingate Sandstone is separated from the underlying Chinle Formation by an erosional disconformity, and the upper contact with the Kayenta Formation is gradational. This causes uncertainty in measurements of thickness, although an average thickness of about 360 feet prevails on Temple Mountain. In most places the Wingate is a light-brown to buff, fine-grained cross-stratified sandstone. The grains are well sorted and contain frosted quartz cemented by calcium carbonate. The origin is probably aeolian. On North Temple Mountain and South Temple Mountain the Wingate is a blue-gray dead-oil sand in many places. Weathered surfaces are oxidized to white. In this area extensive dolomitization has taken place, and some uranium-vanadium ore has been found.

The Wingate-Chinle section exposed on the west end of North Temple Mountain is shown in Figure 1 of Plate 2. The road in the foreground cuts across the contact of the Moenkopi Formation with the Monitor Butte. Extending upward from the two ore chutes are Moss Back strata. Between the Moss Back (light gray) and the mass of Wingate Sandstone above are strata of the upper Chinle Formation. The block of Wingate forming the top of North Temple Mountain appears to expose the entire thickness of the Wingate Formation and possibly a small amount of Kayenta strata at the crest.

Kayenta Formation.—A small thickness of the Kayenta Formation may remain on the extreme top of North Temple Mountain, but the exposed cliffs are inaccessible, and the brownish rock could be a dolomite zone. Typical Kayenta consists of about 280 feet of grayish-brown fine-grained shaly sandstone in lenses 5–10 feet thick. Scattered lenses of mudstone and conglomerate indicate a partial fluviatile origin.

Navajo Sandstone.—The Navajo Sandstone at the top of the Glen Canyon Group is eroded from Temple Mountain and remains only along the reef to the south and east of the mining area. It is unaltered and consists of more than 400 feet of a light-tan, well-sorted aeolian sandstone. Long sweeping cross-bedding is present between horizontal beds as much as 50 feet apart.

Temple Mountain Collapse

The largest concentration of collapse features in the swell occurs at Temple Mountain. Collapsed strata are at the base of the west slope (Fig. 7) in a ridge of bleached and broken Moss Back strata bordered by Moenkopi; this feature is referred to by miners as the flopover. At a higher level on the ridge is a massive rounded slumped block of Wingate Sandstone, sugar loaf (Pl. 1, fig. 2; Fig. 8). The connecting ridge between North Temple Mountain and South Temple Mountain has been subjected to subsidence but not slumping.

The east slope of Temple Mountain (Fig. 9) exhibits a large collapse feature. A somewhat circular area is concentrated in the rough terrain between North Temple Mountain and South Temple Mountain surrounding the candle.

North Temple Mountain and South Temple Mountain form isolated erosional remnants with the main Wingate mass three-quarters of a mile east (Fig. 2). The latter forms one segment

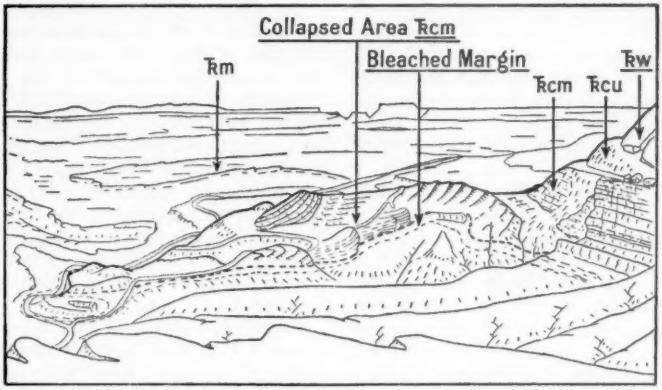

FIGURE 7.—SKETCH OF FLOPOVER LOOKING NORTHWARD FROM SOUTH TEMPLE MOUNTAIN

Slumped blocks of Moss Back (Trcm) and Wingate (Trw) may be recognized in the collapse feature. The main surrounding formation is Moenkopi (Trm), Moss Back (Trcm), upper Chinle (Trcu), and Wingate (Trw) are on the slope at the right.

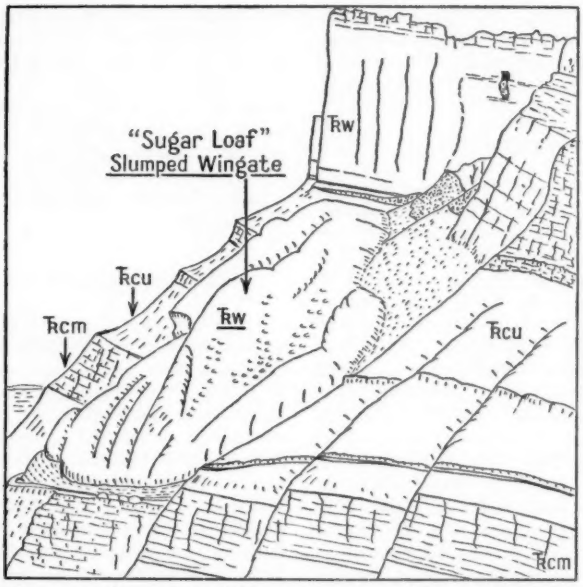

FIGURE 8.—SUGAR LOAF, A DISPLACED BLOCK OF WINGATE SANDSTONE

A sketch looking northward. An adit containing small uranium ore exposures is shown in the upper right of the view. (*See* Figure 7 for identification of symbols.)

of the almost continuous escarpment or reef which encircles the San Rafael Swell.

Encircling Temple Mountain (Fig. 10) is a collar of Moss Back which in places is 100 feet thick and has long been mined for uranium. Between the mountains and the reef is a broad bench underlain by Moss Back and Upper Chinle which has been one of the productive plateau uranium districts (Fig. 2).

The origin of the sunken blocks on the mountain slopes has long been a problem. Casual inspection suggests a landslide, yet this does not

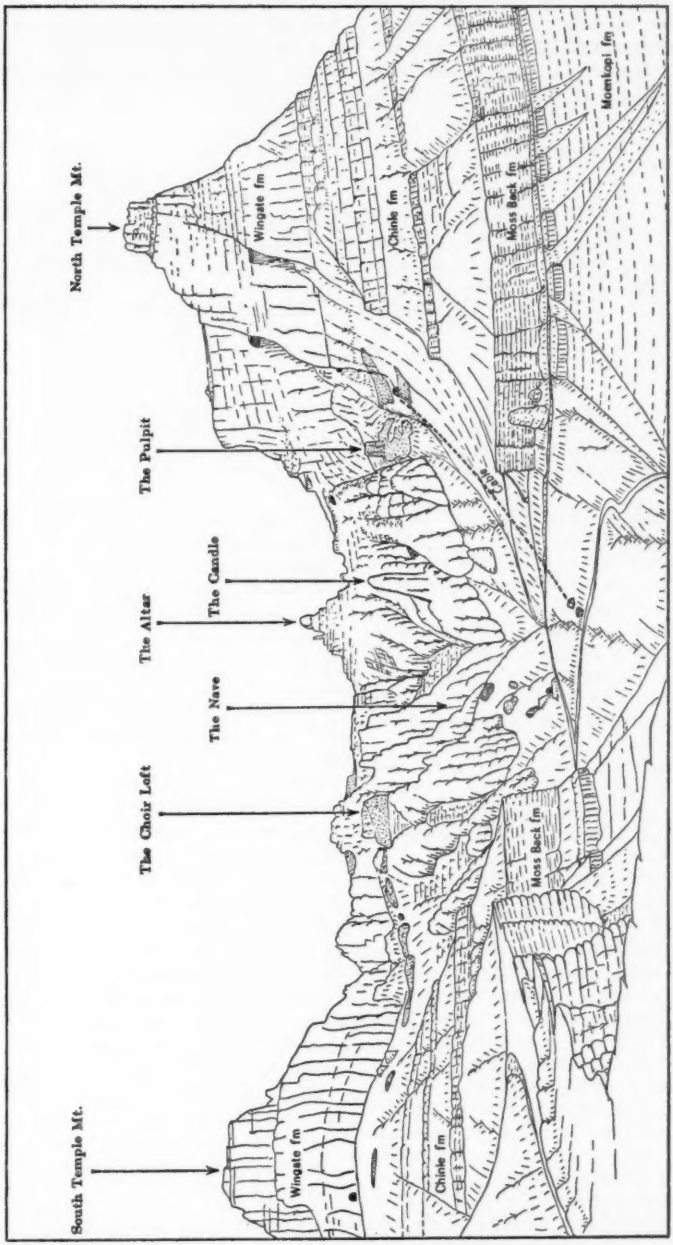

FIGURE 9.—SKETCH OF COLLAPSE AREA ON EAST SLOPE OF TEMPLE MOUNTAIN
The collapse lies between the pulpit and the choir loft.

account for blocks of Wingate hundreds of feet across slumped on either side of the central ridge, or large blocks of Moss Back randomly distributed to the west. Drilling demonstrates

The disturbed array of bleached blocks is attributed to the removal of calcareous material from below, which allowed masses of overlying formations to slump into the cavern created.

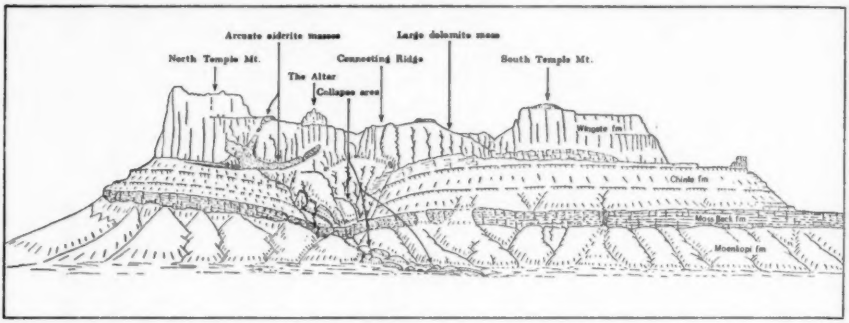

FIGURE 10.—SKETCH OF WEST FACE OF TEMPLE MOUNTAIN

The flopover appears at the base of the mountain with a continuation of the collapse area above. The Calyx area lies on the right.

that these features are deep-seated and lack the superficial character of landslides. Large blocks are in the collapse areas, in contrast to much smaller boulders on talus slopes. In collapse features strata elsewhere resistant to erosion are found in broken segments below their normal stratigraphic position, whereas slip zones along which blocks moved do not appear. The areas of displaced sediments form elevations rather than depressions. The localization of the disturbance between the two peaks and an absence of the disturbance to the north and south where slopes are equally steep is incompatible with normal landslide phenomena. Finally, various collapse areas occur remote from cliffs.

A number of features occur which are not normally associated with landslides. Around the margins of the disturbed zone are concentrations of gossan caused by oxidation of an associated sideritic mineralization. The rocks of the disturbed area are strongly bleached in contrast to an absence of bleaching away from the area. The confused assortment of blocks lies in a structurally depressed area, as shown in the attitudes of surrounding strata. In places veinlets, pods, and streamers of uranium ores occur in the disturbed area and in bordering rocks. A bleached border rock surrounds the broken zone, where the adjacent strata are still intact, although they are in many places bent downward. Dolomitization occurs near the collapsed area and in places forms extensive exposures of dolomitized Wingate Sandstone.

Strata dissolved would be the calcareous limestones of the Sinbad and Kaibab formations. Considerable calcareous soft shale, with some intercalated limestone strata, lies between the Sinbad and Kaibab. Normal Moenkopi strata are also calcareous. A total section of about 150 feet is about one-third calcareous rock. It is possible that solution of calcareous sediment in the region underlying Temple Mountain could have provided adequate space for the collapse at the surface.

The solutions responsible for the collapse were probably hydrothermal. Cumulative criteria support this conclusion: (1) the halo of bleaching which surrounds the collapse; (2) the extensive pyritization associated with the uranium ores; (3) the presence of sphalerite and copper sulfides in close association with uranium in the Wingate Formation high above the ordinary ore-bearing level; (4) the discovery of native arsenic, sphalerite, and hard uranoorganic ore in the west collapse; (5) the occurrence of additional collapse areas in a belt across the swell[1] (Fig. 4).

The Temple Mountain collapse area contains the flopover, sugar loaf, the nave, and perhaps two arcuate intermediate areas along the ridge in which mineralization is apparent, but collapse is a minor feature.

[1] Several areas lie west of the region shown in Figure 4.

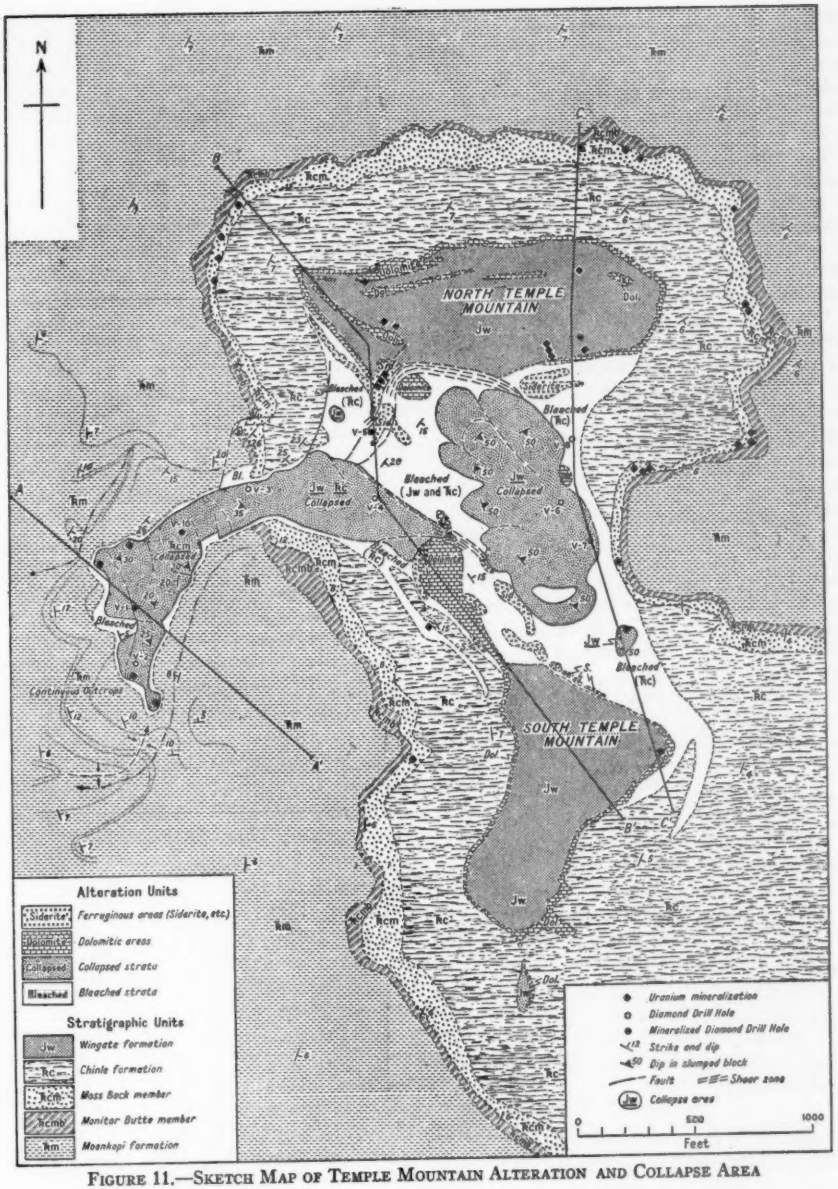

FIGURE 11.—SKETCH MAP OF TEMPLE MOUNTAIN ALTERATION AND COLLAPSE AREA

Flopover

A low ridge (Fig. 7) irregular in slope and strewn with displaced blocks of Moss Back strata marks the collapse west of Temple Mountain. It is bordered on the north, west, and south by the Moenkopi Formation and abuts sugar loaf on the east. The flopover is

1400 feet long and is structurally incompatible with moderately inclined mesas and platforms of red and brown Moenkopi which flank it on either side. The ridge slopes westward from Temple Mountain at about 9°. The trend is S. 25° W. for 900 feet and south for 500 feet.

The tongue lies in a basinlike structural trough where the Moenkopi strata are inclined at angles of a few degrees to as much as 23° toward the ridge. The ridge represents the locus of a synclinal trough in which the Moenkopi strata on either side deviate from the regional dip and incline toward it.

Bordering the flopover is a narrow, light-brown band, ranging in width from 3 to 50 feet, in which Moenkopi beds have been altered to yellowish brown and deformed but are continuous with near-by strata. In the core of the collapse, deformation has broken all continuation with the Moenkopi on either side, and bleaching is intense. Lenses of conglomerate, coarse granular texture, bedding, and a few logs identify the bleached strata within the collapse with the Moss Back Formation. The Moss Back, however, is distributed in structurally distinct blocks, some more than 150 feet across and as much as 50 feet thick. In general, the blocks are inclined northwestward at angles of 20°–25°. On the geologic map (Fig. 11), block attitudes are indicated by special dip symbols.

The nearest Moss Back strata in place lie below sugar loaf at the east end. Here strata about 75 feet thick are inclined eastward at 13°.

Urano-organic ores occur at several isolated points on the flopover on the surface and in drill holes. One of the surface exposures on the Fumarole claim has produced ore. A few minor occurrences of similar material have been noted elsewhere.

Sugar Loaf

A rounded block of Wingate Sandstone (Pl. 1, fig. 2; Fig. 8), sugar loaf, forms a continuous exposure about 400 feet in diameter, adjoining the flopover. The block is bordered north and south by highly bleached sandstones and shales of the Chinle Formation. The block (Figs. 8, 10) lies about 50 feet below the base of the Wingate which caps North Temple Mountain and the ridge between North Temple Mountain and South Temple Mountain.

Sugar loaf represents a large block of Wingate Sandstone. The Chinle below has slumped into its present position. The amount of displacement from a likely original position would involve a downward vertical component of at least 300 feet.

The alteration, deformation, and contortion of the Chinle Formation surrounding the block are significant. In an irregular halo which ranges from a few feet to 200 or more feet in width, the Chinle has changed from a normal

PLATE 1.—VIEWS OF TEMPLE MOUNTAIN

FIGURE 1.—West slope of Temple Mountain. White bleached area in the center of the slope is the flopover. Black arcuate area on North Temple Mountain is mainly siderite.

FIGURE 2.—Air view of North Temple Mountain looking east. A dislocated block of Wingate Sandstone is shown midway on the west slope. A portion of the main mining area, known as the Calyx area, appears in the upper right corner.

PLATE 2.—STRATIFIED ROCKS ON TEMPLE MOUNTAIN

FIGURE 1.—Wingate Sandstone crest of North Temple Mountain. West slope is shown in the foreground. Light-gray unit above the road is Moss Back (Chinle). Dark strata between the Moss Back and Wingate crest are Chinle. Dark strata at the bottom are Moenkopi. The Monitor Butte forms a poorly exposed thin dark zone at the base of the Moss Back.

FIGURE 2.—Dolomitized Wingate Sandstone. Dark-gray massive rock in the center of the view is dolomite. The nearly horizontal strata below are upper Chinle. Below the dolomite the Chinle strata have been bleached almost white.

PLATE 3.—FEATURES OF THE SAN RAFAEL SWELL

FIGURE 1.—Collapse no. 4, looking south toward Temple Mountain. Low light-gray rounded hills in foreground form part of collapse no. 4. The bench east of Temple Mountain is the site of the main mining area.

FIGURE 2.—Tar seep near collapse no. 4. Viscous nonradioactive tar seeps out of Sinbad Limestone strata. The two pencils indicate size of seep.

FIGURE 3.—Sink hole in Kaibab Limestone, San Rafael Swell. No apparent alteration is associated with this sink hole which appears to be a normal solution collapse. It is located on the edge of a valley in the northern part of the Swell.

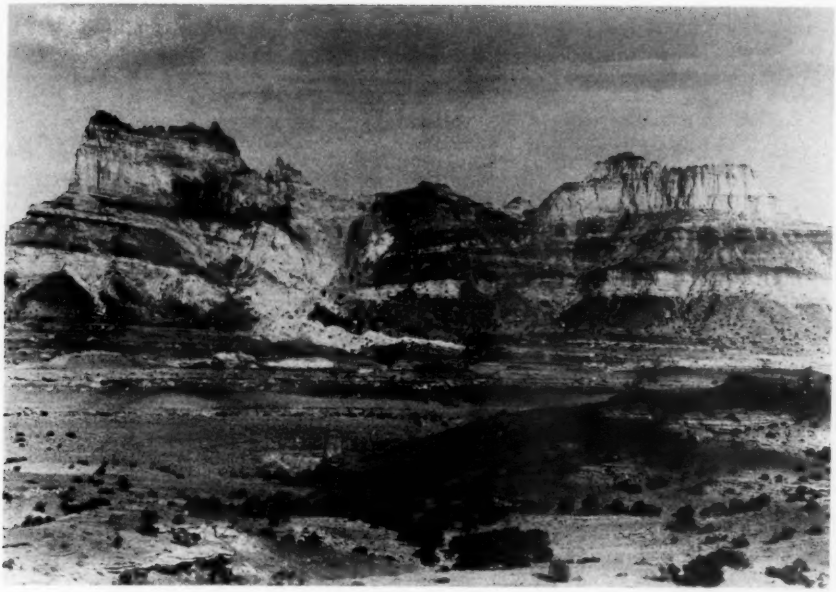

FIGURE 1

FIGURE 2

VIEWS OF TEMPLE MOUNTAIN

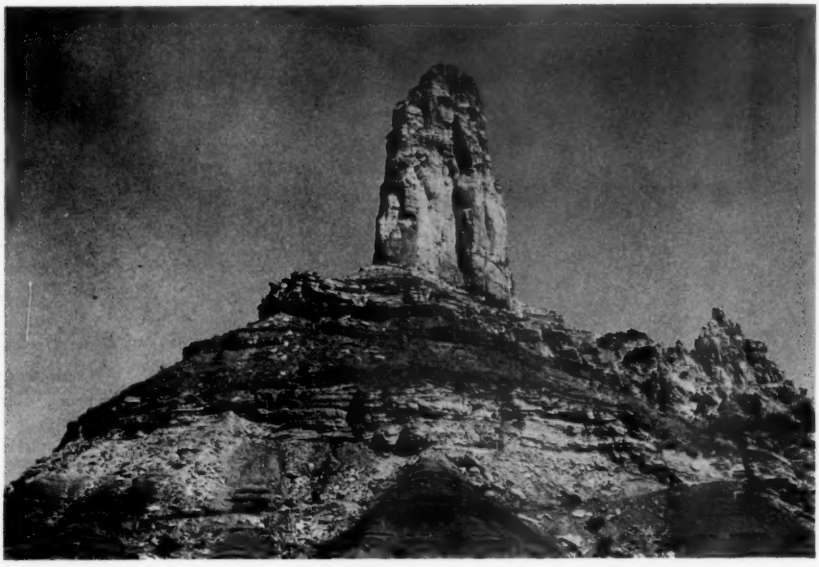

FIGURE 1

FIGURE 2

STRATIFIED ROCKS ON TEMPLE MOUNTAIN

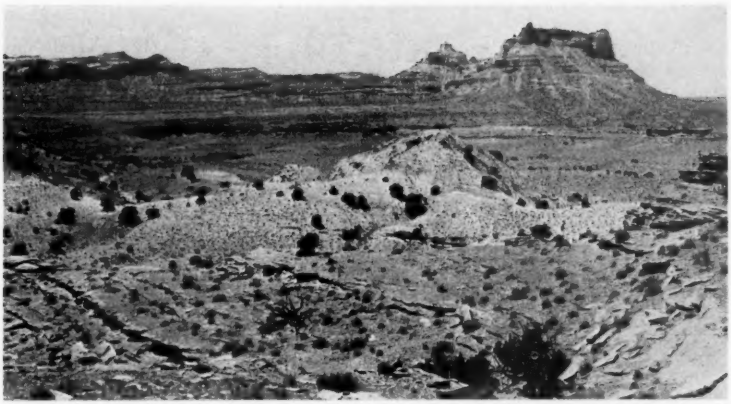

FIGURE 1

FIGURE 2

FIGURE 3

FEATURES OF THE SAN RAFAEL SWELL

deep red or brown to white, light gray, or greenish gray. More porous beds show more intense bleaching which in places extends in streaks considerably beyond the general boundary of the halo. Chromium mica clay has also precipitated along fractures. The bleaching is attributed to hydrothermal solutions.

The deformation and contortion of the Chinle surrounding the downdropped block may provide a clue to the mechanism of emplacement. In the vicinity (Fig. 11) two conspicuous local marker beds in the Chinle deviate in strike about 45°: a gray conglomerate, coarse sandstone, and grit about 70 feet above the Moss Back and a red sandstone 30 feet higher which contains rosettes of calcite. The dip of the marker beds changes from a normal angle of 7° eastward to as much as 30° south. As the dip increases the two beds squeeze to within 25 feet of each other and terminate against the Wingate Sandstone. Where these beds are cut off, highly altered and contorted sandstones and shales of the Chinle continue to encircle sugar loaf. In one place these form a zone not more than 10 feet thick between the lower edge of the block and a mass of Moss Back Sandstone.

Slumping necessary to put sugar loaf in its present position probably occurred while the strata of the San Rafael Swell were covered with Wingate strata over the central dome. Hydrothermal solutions produced underground stoping into which the Wingate block slumped. Later, during the late Tertiary erosion interval, the Wingate block became a ridge, since it was more resistant to the agencies of erosion than the surrounding Chinle. Thus a filled depression is reflected in the topography as an elevation.

Ridge

The ridge, underlain by Wingate strata, extends for about 1500 feet between North Temple Mountain and South Temple Mountain. It forms a long, narrow bridge, partly resistant to deformation, between the collapse areas on the west and east.

The ridge contains the most highly dolomitized portion of the Wingate observed. On the west facade, extending northward about 800 feet from South Temple Mountain, the Wingate strata have been altered to a dark, dull-brown dolomite (Pl. 2, fig. 2). Numerous vugs or open cavities have developed which are lined with myriads of small rhombohedral dolomite crystals. In many places the original granular texture of the Wingate Sandstone has been obliterated. The dead-oil sand characteristic of the core of the ridge as well as North Temple Mountain and South Temple Mountain has disappeared. However, a nonradioactive asphaltite has developed in the cavity openings and forms nodules which rest upon rhombohedral crystals.

The asphaltite is normally brittle but may deform on extremely warm days. It is less fluid than the viscous tar along fractures in the mining area and not so hard and brittle as the asphaltite in siderite areas. None of these three asphalt occurrences is ore-bearing. However, on Temple Mountain in Wingate and Moss Back strata, hard brittle urano-organic impregnations of ore grade occur in sandstone. About 12 of these occurrences have been found in the Wingate Formation, whereas many hundreds of feet of mine workings encircle North Temple Mountain and South Temple Mountain in the Moss Back.

The strata of the ridge form a low but somewhat angular synclinal sag (Fig. 10). However, the structure is cut by shear zones in several places and has suffered fracturing near the collapse areas on either side. Several shear zones that cut the ridge are arcuate in trend and have a configuration indicated by curved ferruginous patches (Pl. 1, fig. 1).

The depression of the ridge (about 150–200 feet) below the Wingate of North Temple Mountain and South Temple Mountain is much less than that of the collapse of the nave and sugar loaf. However, the effects of bleaching, dolomitization, and ferrugination are more pronounced. The ridge has been affected by the solutions associated with the collapse phenomena; it may represent a structurally less active but thermally more active intermediate area.

Nave

An area of collapse about 1000 feet north-south by 600 feet east-west lies east of the connecting ridge (Fig. 9). The area ranges in elevation from about 5900 feet in the canyon on the eastern margin to 6500 feet along the ridge. The nave is cut by a series of radial drainage lines consisting chiefly of small angular canyons, carved in bedrock or talus, which have gradients of 30°–50°. These converge just below the band of the Moss Back Formation that cuts across the eastern margin.

Encircling the nave on the north, west, and south is a semicircular rampart of the Wingate Formation. The rampart embraces parts of

North Temple Mountain, the connecting ridge, and South Temple Mountain.

The Chinle Formation is strongly bleached within the area of the collapse and for some distance on either side (Fig. 11). In most places the Chinle red and brown is bleached to a pale greenish gray or white. Strata are warped but continue until broken at the core of the collapse. Near the top of the Chinle Formation and extending into the Wingate highly ferruginous concentrations occur as bands and lenses.

The collapse core forms a circular area about 600 feet in diameter, bounded on the north by the pulpit and on the south by the choir loft (Fig. 9). In the core a large mass of the Chinle Formation has been removed by solution, and into the vacant space great blocks of the Wingate Formation have slumped. Wingate blocks now lie at an inclination of about 50°; each slopes toward the center of the collapse. Some blocks are 100 or more feet thick; in many places around the margins they lie unconformably upon bleached Chinle strata almost at right angles to the stratification yet are firmly bonded at the surface of contact. On the south (Fig. 11) one isolated slumped Wingate block has been penetrated at the base by a small adit in which are exposed firmly cemented breccia blocks. The breccia contains uranoorganic ore in fragments and matrix.

The evidence of bleaching is important to an explanation of the origin of the nave collapse. The upper part of the Chinle Formation is highly altered for 1400 feet southward from the margin of the collapse core, and alteration is accompanied by ferruginous concentrations and masses of dolomitization. Clays have changed from red or brown to pale greenish gray. Sandstones are light gray or white. Close to the collapse, for 200 feet above the Moss Back on both sides, most of the underlying rocks are concealed by talus. However, the light color of the soil and a few exposures indicate that most of the talus-covered area is bleached. The extent of alteration in the Moss Back is difficult to ascertain because of surface staining, but this porous member may have been affected for a considerable distance.

Apparently thermal solutions accompanying dolomitization and responsible for extensive alteration were present in sufficient abundance to cause destruction of underlying carbonate rock constituents. This destruction caused natural-gravity stoping. The blocks settled gradually with little break other than a shear zone to separate the moderately inclined strata from the slumped strata. Subsequent erosion yielded the present nave topography.

Subsurface Interpretation

The structure of the collapsed area at Temple Mountain is illustrated by sections A-A′, B-B′, and C-C′ (Figs. 11, 12, 13, 14) constructed from surface and drill-hole data. In general, the sections indicate a floor of Coconino Sandstone beneath the collapse features. In cores the granular character and regular stratification of the Coconino are retained, and no evidence of brecciation is observed. The level at which the Coconino was encountered in each drill hole agrees with a projection of the regional dip of the formation, allowing for minor offsets. The maximum dislocation inferred is suggested in section B-B′ near drill hole V-5 (Fig. 13). Here the displacement of the Coconino might reach 75 feet, although a considerably smaller displacement is likely.

Contours indicating the approximate upper surface of the Coconino, as inferred from drill data, are shown in Figure 15*a*. Faults of minor displacement are postulated. These are based on the regional fault characteristics and account for the difference between the geologically projected level and the level encountered by drilling. In Figure 15*b* contours are corrected for regional dip, and faulting has been eliminated. The result is a small sag or basin structure, within the Coconino, which extends beyond the flopover and the nave. The absence of a major sag indicates relatively little displacement of the Coconino in contrast to the displacement in the overlying strata. The existence of some sag, however, shows that the structural aspect of the collapse phenomenon extends in depth at least into the upper part of the Coconino. Since the sag agrees in alignment with the collapse, displacement is presumably contemporaneous. Thus it is as young as the youngest strata affected by the collapse. It is at least post-Wingate and could be much later.

Where the Coconino Formation is exposed on the San Rafael Swell, eastward-trending normal faults cut it in many places with displacements which range from a few feet to more than 100 feet. Such faulting probably lies beneath the Temple Mountain collapse. Although faulting is not observed on the surface, it appears on line of strike in the Vanadium King workings to the east.

The Coconino has been mineralized by galena in scattered spots which are surrounded by oxidized material. In places, it is impregnated

with viscous or oxidized petroliferous matter; in one place it contains urano-organic ore. Diamond-drill cores from the collapse zone above the Coconino Formation where Kaibab

The sunken blocks of broken and deformed strata of the Temple Mountain collapse rest on a moderately tilted platform of Coconino strata. Broken sedimentary strata moved down-

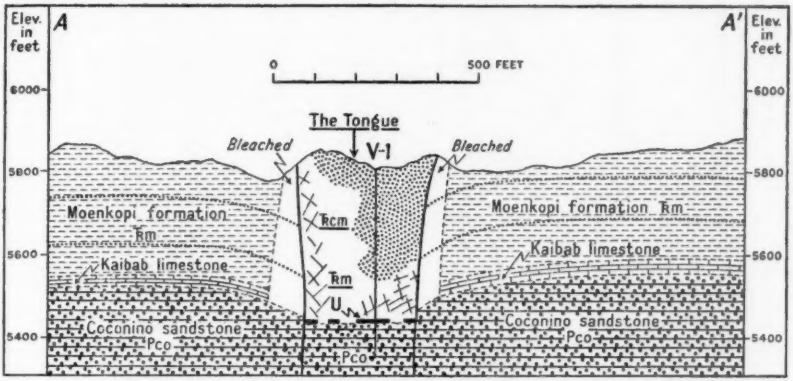

FIGURE 12.—SECTION A-A' THROUGH TEMPLE MOUNTAIN COLLAPSE
See Figure 11 for location of section.

strata would be expected are highly brecciated and are characterized by angular fragments apparently of broken Kaibab strata. Thin sections of cores exhibit clay-mineral alteration and demonstrate the removal of carbonates. In places sulfides, including galena, pyrite, and sphalerite, have been deposited. Some urano-organic ores have also accumulated in this zone and in the overlying broken material. The Kaibab appears to have been destroyed, and only breccia remains.

Brecciation extends upward 200 feet in places and includes a large volume of Moenkopi strata. The latter have suffered alteration similar in many respects to the breakdown of the Kaibab.

Large blocks of Wingate, Chinle, Moss Back, and Moenkopi strata may be recognized in portions of diamond-drill cores from the collapse area. The boundaries between blocks, however, are not delineated everywhere, and in many instances individual core fragments are not recognizable stratigraphically.

Cores of porous sands throughout the collapse area are impregnated with petroliferous matter. The material is pungent and for the most part not indurated. Although it is viscous in only a few places, it appears, for the most part, to represent petroliferous, nonuranium-bearing impregnation. Hard, brittle urano-organic ores are also encountered, however, and numerous instances of uranium mineralization in such material have been observed.

ward because of gravity, and solutions provided upward chemical transport, alteration, and deposition.

The resultant form is a section of collapse and brecciation curved in outline which rests on a tilted and slightly displaced platform of Coconino Sandstone. The border area consists of altered and deformed but not displaced strata.

The downwarp of the Moss Back is shown by structure contours on the base of this formation (Fig. 16). The warped surface is broken by the collapsed areas. The base of the Moss Back near the collapse has been somewhat warped because of structural influence.

OTHER COLLAPSE AREAS

Distribution

The collapse mechanism has been noted at 10 areas in the southern part of the San Rafael Swell in addition to Temple Mountain. Excepting 3 which lie to the west these are numbered consecutively in Figure 4. All are surrounded by strata which are older than the Wingate Sandstone of Temple Mountain. In most, the surrounding formation is Moenkopi, although a stratigraphic spread of at least 500 feet exists between the exposed surfaces of collapse No. 1 near the top of the formation and collapse no. 9 at the base.

With the possible exception of nos. 9 and 10

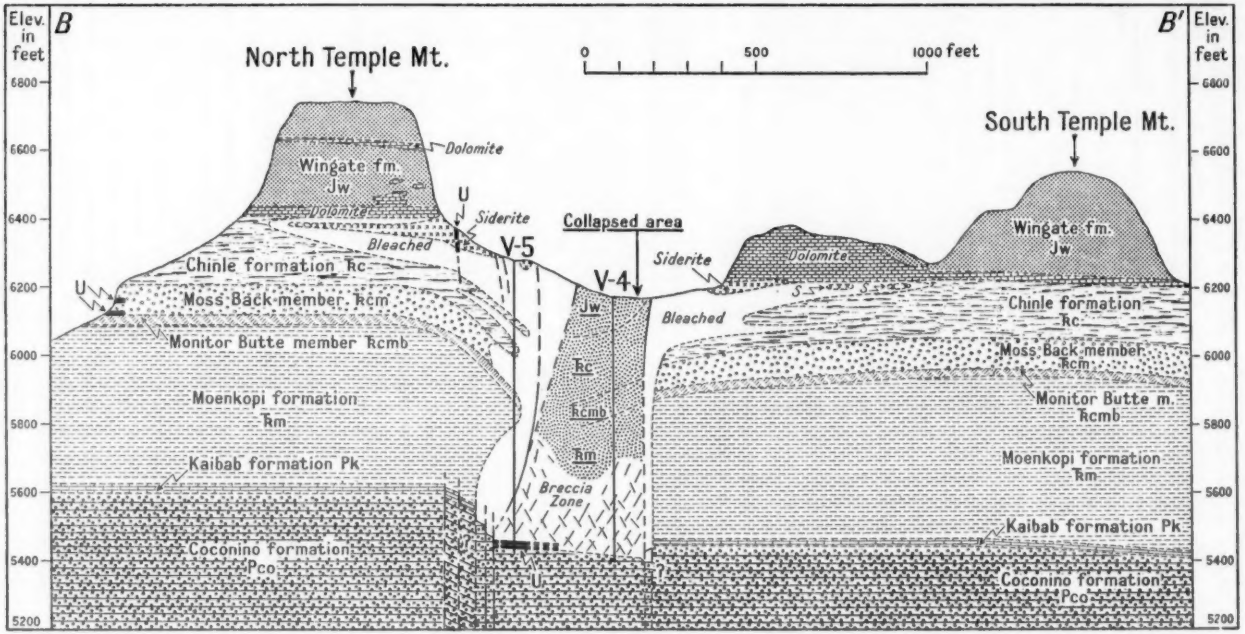

FIGURE 13.—SECTION B-B' THROUGH TEMPLE MOUNTAIN COLLAPSE
See Figure 11 for location of section.

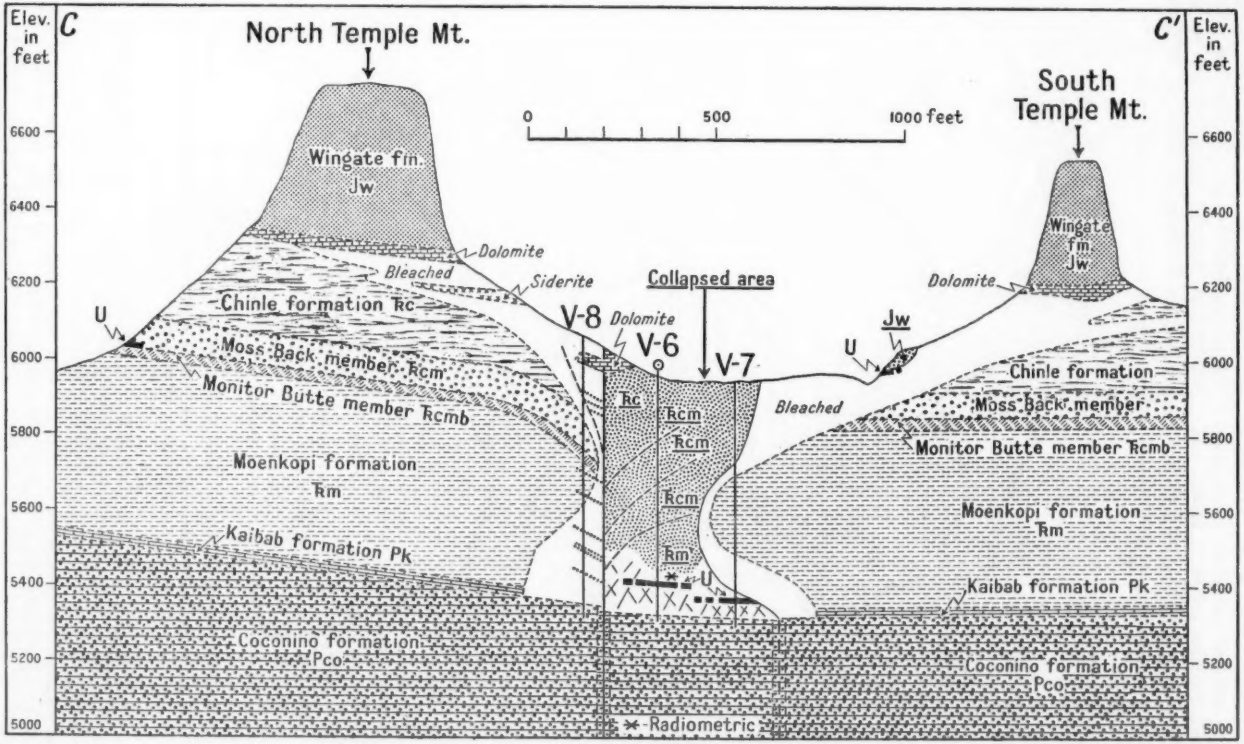

FIGURE 14.—SECTION C-C' THROUGH TEMPLE MOUNTAIN COLLAPSE
See Figure 11 for location of section.

certain features are common to all these collapse areas. Each collapse occurs in a structural trough. Each contains a core of broken sediment dropped from overlying strata which are easily recognized in the argillaceous phases. Many sandstones are impregnated with petroliferous material which oxidizes on the surface and produces a white or light-gray

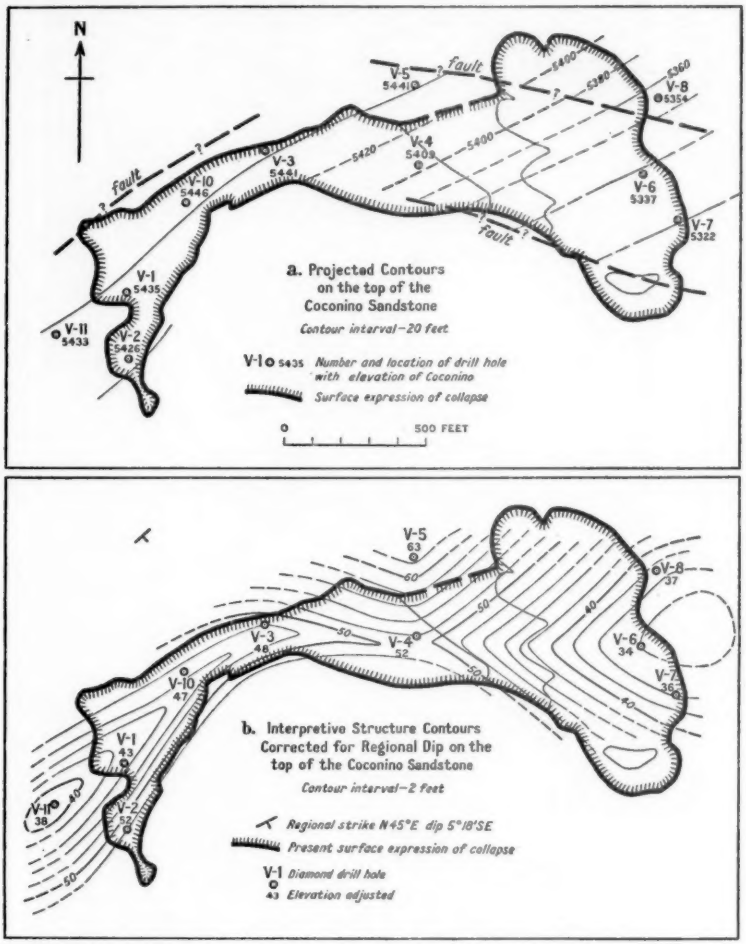

FIGURE 15.—PROJECTED CONTOURS ON THE TOP OF THE COCONINO SANDSTONE BENEATH TEMPLE MOUNTAIN

no longer adjacent. Each shows two zones of alteration: One zone forms an outer marginal band which coincides with warped but unbroken strata and encircles the broken core. The other consists of the more highly altered broken rock of the core.

At the collapse areas alteration is more surface coating. The latter should not be confused with the color change which accompanies alteration and results in the removal of iron and readjustment of the clay-mineral assemblage.

The area designated as collapse no. 9 lies at the base of the Moenkopi Formation and in the

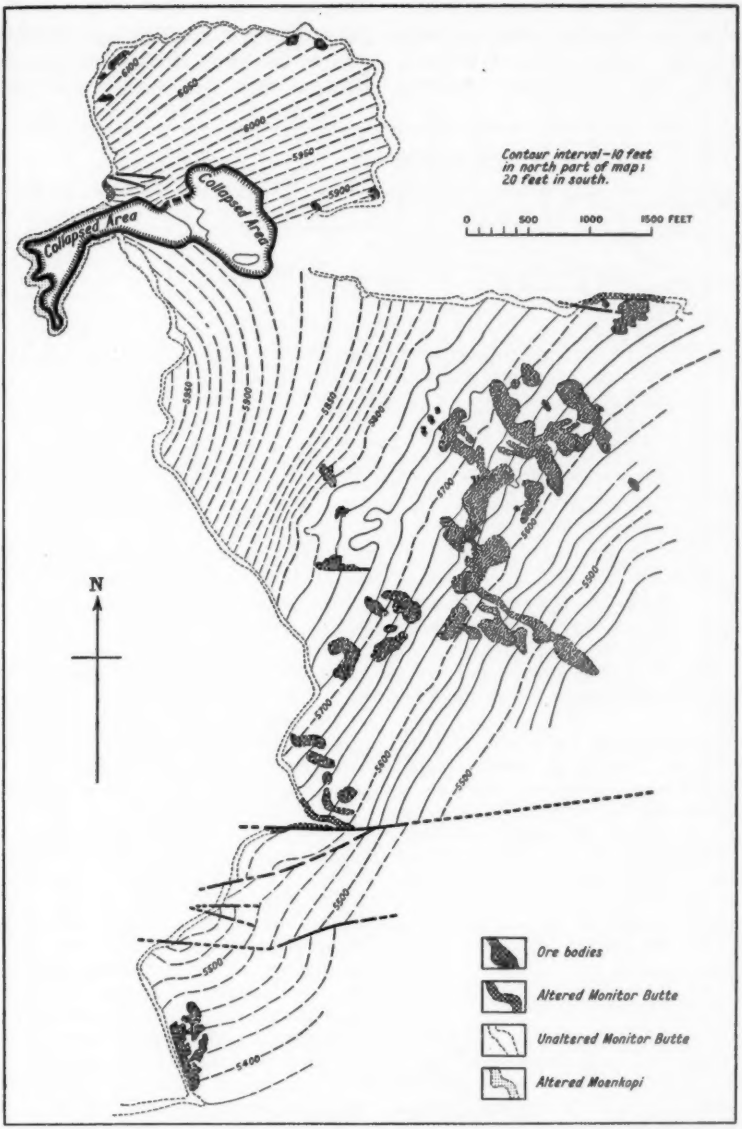

FIGURE 16.—PROJECTED CONTOURS ON THE BASE OF THE MOSS BACK SANDSTONE, TEMPLE MOUNTAIN

Kaibab horizon just above the Coconino. Because of the depth to which it has been eroded it appears to expose faulting which may lie below the surface at other collapse features less deeply eroded. At Temple Mountain and at each of the other collapse features, no connection that can be traced on the surface exists between faults and the structural sag which

surrounds the collapse. At collapse no. 9, however, faulting cuts the altered area that marks the core of the collapse. Because of the proximity of the Coconino platform below, the upper edge of the collapse is 70 feet below the top of the Moenkopi. Stratigraphically at the lowest elevation in the collapse, blocks of Moss Back Sandstone lie 240 feet

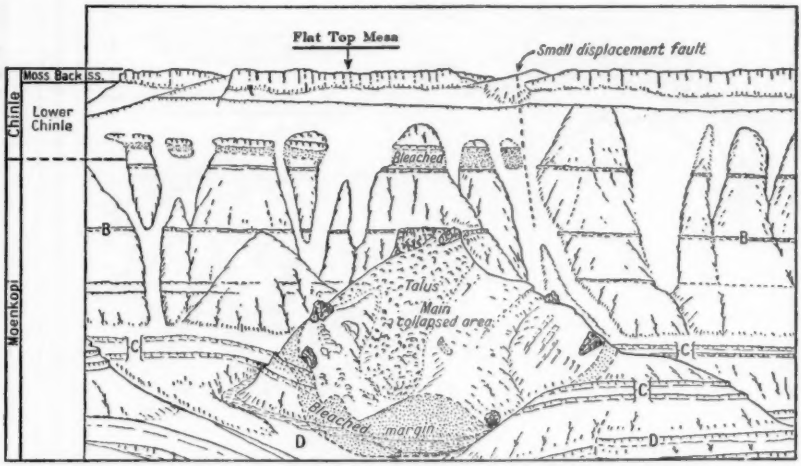

FIGURE 17.—SKETCH OF EASTERN COLLAPSE ZONE OF COLLAPSE NO. 1
B, C, and D zones correspond to Figure 19.

accumulation of debris in the collapse core is less than it is in the collapse features higher in the section.

Uranium ore has been found only in association with the Temple Mountain collapse. Scintillometer surveys failed to disclose abnormal radioactivity at other collapse areas. Drilling in the collapse cores at 2, 4, and 8 also failed to disclose associated ore bodies. Although the holes at 2 and 4 were drilled with a churn drill and the cuttings were not available, the results indicated that no ore was discovered. The diamond-drill core for 8 showed alteration but no mineralization.

Individual Areas

COLLAPSE NO. 1: Collapse no. 1 (Fig. 4) lies west of Flattop Mesa (shown on some maps as Shinarump Mesa). A collapse mound (Fig. 17) is located at one end of an elliptical structural depression about 700 feet long. The depression contains two collapse zones; each is roughly circular and has an eastward alignment (Fig. 18). Broken blocks of Moss Back Sandstone are prominent in the core of the collapse, but the nearest exposure in place is on the top of Flattop Mesa, 420 feet east and 132 feet above. below their projected stratigraphic position (Fig. 19).

The structural depression is surrounded by red siltstones, mudstones, and interbedded buff sandstones of the upper Moenkopi; these have a regional inclination of 2°–3° to the southeast. The change in attitude from the general structural slope to the local depression is gradual. The beds are warped in a circular arrangement and have dips of 3°–7° toward the depression center. Near the core dips increase to 8°–14°, and the trough reflects an eastward elongation. However, unbroken beds of the Moenkopi nowhere exceed 14° in dip. Interpretations of the two collapse zones and accompanying structure are shown in Figures 20 and 21.

The larger collapse zone is characterized by bleached, recemented, and poorly sorted Moss Back sandstone blocks and fragments and minor amounts of lower Chinle siltstones and sandstones. Few, if any, Moenkopi blocks are involved. Because of variable resistance to erosion the collapsed core material exhibits irregular topography. The stratified collapsed blocks are inclined to the west; strike directions of 12 blocks are oriented in a quadrant within an arc N. 40° W. to N. 50° E. The amount of

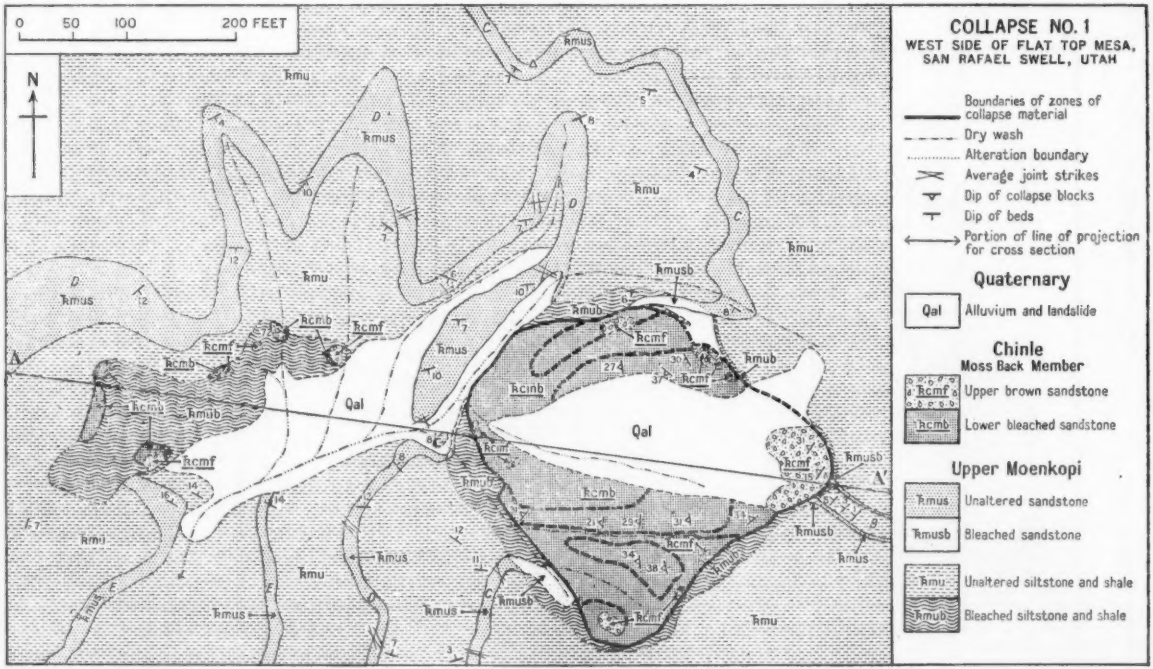

FIGURE 18.—ALTERATION MAP OF COLLAPSE NO. 1
B, C, D, and E zones correspond to Figure 19.

dip ranges from 15° to 38°. The orientation of collapsed material is a noticeable feature here and in other collapse cores.

from deep red or buff to a pale yellowish brown within a few feet of the core. The color change has been mapped. It is about 5–15 feet wide

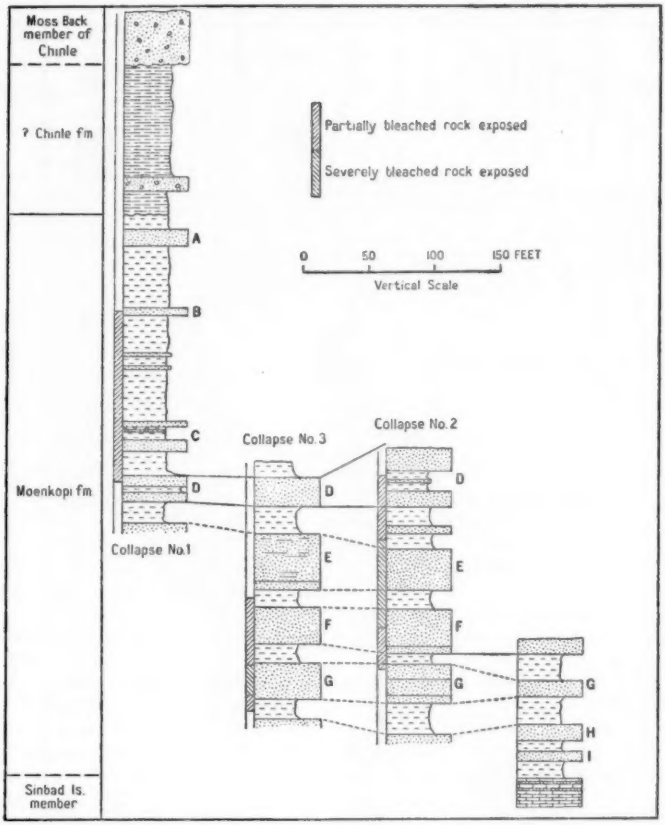

FIGURE 19.—STRATIGRAPHIC CORRELATION BETWEEN COLLAPSES NOS. 1, 2, AND 3 AND THEIR RELATION TO THE SINBAD MEMBER OF THE MOENKOPI FORMATION

The western collapse zone is poorly defined and has little accumulation of core material. Seven patches of Moss Back Sandstone, which border Moenkopi siltstones and mudstones bleached *in situ*, average up to 25 feet in diameter. The ground between the east and west collapse zones (Fig. 18) is partly covered by alluvium. The isolated patches of Moss Back Sandstone either mark the edge of the collapse or remnants of a cover once above most of the larger depression.

Shale and sandstone of the Moenkopi change adjacent to the eastern collapse and probably represents bleaching due to solution and reduction of iron.

Alteration effects appear within the eastern core (Fig. 18). In places grit and conglomerate are silicified, but the time of silicification is uncertain. Several small carbonate veins also occur in the sandstone. The most impressive feature, however, is the bleaching of the sandstones, shales, and clays, which produces a characteristic white.

The center of the smaller western collapse is

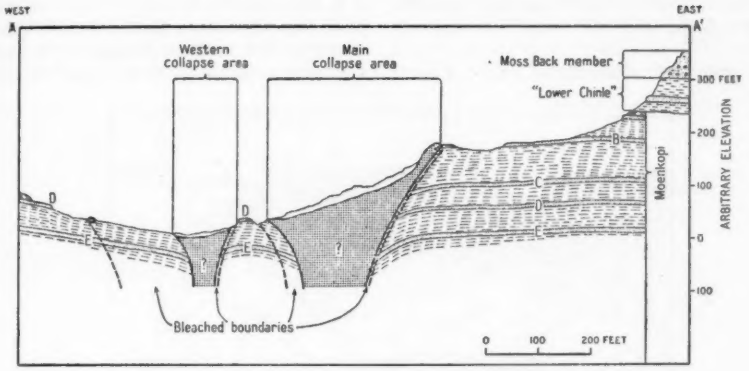

FIGURE 20.—SCHEMATIC CROSS SECTION OF COLLAPSE NO. 1
See Figure 18 for location of section. B, C, D, and E zones correspond to Figure 19.

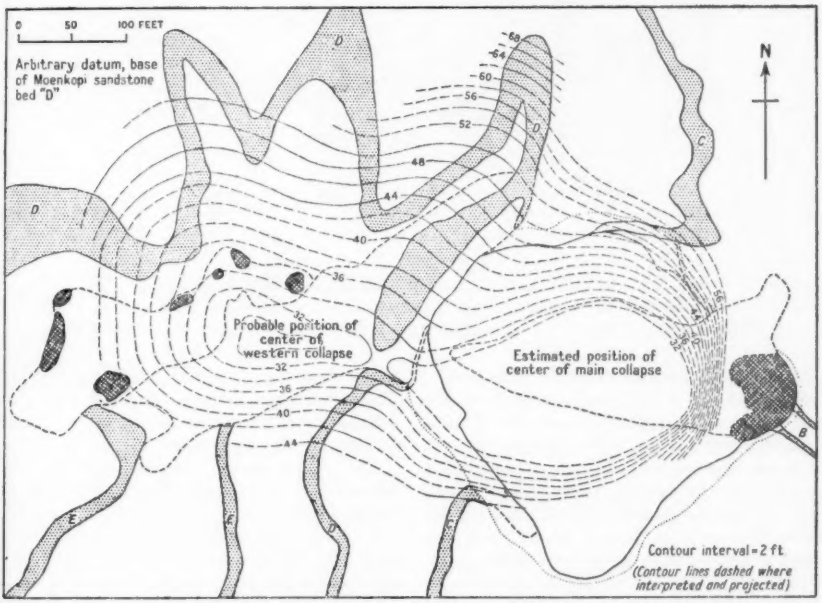
0 50 100 FEET
Arbitrary datum, base of Moenkopi sandstone bed "D"
N
Probable position of center of western collapse
Estimated position of center of main collapse
Contour interval = 2 ft
(Contour lines dashed where interpreted and projected)

FIGURE 21.—STRUCTURAL INTERPRETATION OF THE CONTOURS OF COLLAPSE NO. 1
B, C, D, and E zones correspond to Figure 19.

obscured by alluvium, and the rim of bleached sediments is less noticeable. An irregular band of bleached siltstones and mudstones 30–105 feet wide borders the northwestern edge. This collapse zone is smaller in plan, the alteration effects are less intense, and the vertical influence of the collapse is considerably less. Mapping indicates that the bleached rims are wide on the west but narrow on the eastern edges.

Small patches of hydrocarbon with a greenish micaceous stain impregnate sandstone near the center of the eastern collapse zone. Partial chemical determinations on a grab sample

from one of these patches reveal less than 0.01 per cent U_3O_8, 0.10 per cent V_2O_5, and no $CaCO_3$.

COLLAPSE NO. 2: Collapse no. 2 lies 1.5 miles southeast of Flattop Mesa and 2 miles southwest of Temple Mountain. It is south of the road between Flattop Mesa and the Temple Mountain camp (Figs. 3, 4). The chaotic core forms an isolated hill which is anomalous in contrast to the normal topography of surrounding Moenkopi beds (Fig. 22). No surface radioactivity above normal background was noted at any place in the collapse area.

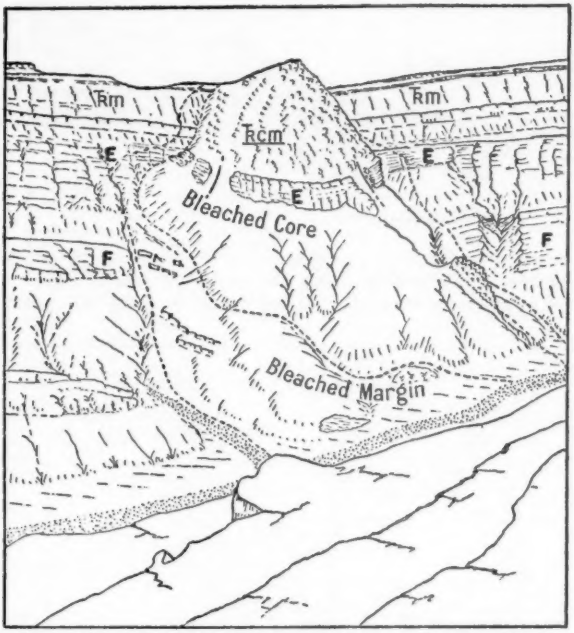

FIGURE 22.—SKETCH OF THE SOUTHERN CORE OF COLLAPSE NO. 2
Looking south in the Temple Mountain area. E and F zones correspond to Figure 19.

The beds cut at the surface by the collapse are the sandstones, siltstones, and shales of the lower Moenkopi (Fig. 19). Alteration is most noticeable in the E and F sandstones and the shales and siltstones between them. These beds are about 100–150 feet above the Sinbad Limestone and about 300–350 feet below the Moss Back.

The collapse lies in an elliptical structural depression approximately 900 feet by 400 feet with the long direction north-south. Of the two collapse centers, the southern is more extensive; it is characterized by an oval chaotic core of Moss Back Sandstone blocks about 100 feet in diameter. The northern collapse is a subsidiary depression outlined by bleaching of the Moenkopi sediments; Moss Back Sandstone is absent. The surrounding Moenkopi beds dip toward the collapse cores. In places immediately adjacent to the core the dips exceed 20°.

The surrounding bleached zone is more altered next to the core. The E bed (Fig. 19) forms a collar around the southern core (Fig. 22). It is normally a red-brown sandstone which gradually changes to tan brown to buff as it approaches the core. In the core the bed is white to light gray and may be dark gray and black if freshly broken. Considerable dead-oil stain is common within the bleached zone. The Moss Back Sandstone, which constitutes the core, has also been severely bleached, and along fractures small solution channels represented by closely spaced tubular streaks about 2 mm in diameter have formed. These seem to have been derived primarily at the expense of solution of the matrix and partial solution of some of the larger detrital grains. The quartz grains,

which form the walls of these channels, are considerably more rounded than in typical Moss Back. This suggests solution of the quartz.

The northern zone has a similar pattern of brick-red to tan shales, siltstones, and sandstones which are bleached to a light tan or dark gray in the collapse. In the core of the collapse, much of which is only partly exposed, the beds

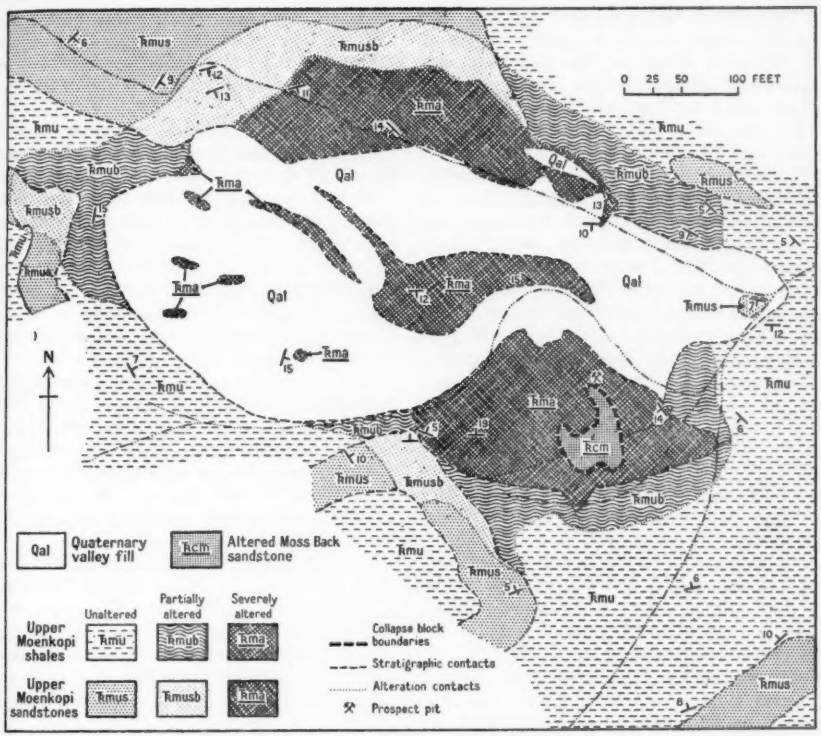

FIGURE 23.—ALTERATION MAP OF COLLAPSE NO. 3

bleaching. The central portion consists of severely bleached but only moderately displaced Moenkopi sediments.

COLLAPSE NO. 3: Collapse no. 3 is located about half a mile east of Flattop Mesa and about 1.5 miles west of Temple Mountain (Figs. 3, 4). It lies in a small canyon, partly hidden by alluvium.

It is elliptical, trends east, and is about 700 feet long and 500 feet wide. The displaced or chaotic core measures 50 by 80 feet, and the Moss Back Sandstone present represents a displacement of about 300 feet.

The rocks bordering the collapse constitute the lower 100 feet of the upper member of the Moenkopi Formation, the F and G sandstone beds (Fig. 19). These consist of interbedded

are bleached to light gray or almost white, but on freshly broken surfaces they are dark gray to black. This zone is about 500 feet by 350 feet. Much of the rock within the core shows traces of petroliferous material. Surrounding the central zone is a slightly bleached marginal zone which ranges from 30 feet to more than 100 feet in width. The alteration map (Fig. 23) of collapse no. 3 indicates the above-mentioned zones as unaltered, partially altered, and extensively altered Moenkopi sediments.

The area is structurally similar to the previously described collapses. A general dip of 5°–10° to the southeast prevails, but the beds change attitude near the collapse core and form a structural basin. On the north and west sides

of the collapse the dips are 15° or more toward the center, but on the west and south the dips rarely exceed 10°. This is to be expected if the regional dip to the southeast is considered. An exception to this is the immediate vicinity of the core, where dips on all sides approach 20°. The cover interferes with a detailed picture of the structure. Probably this represents one collapse zone rather than a line of collapses as at collapse no. 1 or Temple Mountain. No attitudes were noticed that indicate more than one central core.

The collapse core lies in an intensely bleached area (Fig. 23) and forms a small hillock of jumbled blocks of Moss Back Sandstone and surrounding bleached Moenkopi beds. Coarse-grained sandstone, some conglomeratic material, and woody trash remains identify the unit as Moss Back Sandstone. The hillock is more than 300 feet vertically below the projected position of the Moss Back Sandstone.

The collapse has been staked as a uranium prospect. A small discovery pit has been dug (Fig. 23), but no indications of unusual radioactivity were noted in a reconnaissance with scintillation equipment. The basinlike depression of the country rock, the jumbled core, and the surrounding halos of alteration conform to other collapse features.

COLLAPSE NO. 4: the largest collapsed area Pl. 3, fig. 1) outside of Temple Mountain, is located on the southeastern flank of the San Rafael Swell, 1.5 miles northeast of Temple Mountain. The surrounding area is underlain by moderately inclined Moenkopi sediments (Figs. 3, 4), which are stratigraphically at about the level of the Sinbad Limestone Member and the lowest part of the upper menber of the Moenkopi Formation (Fig. 19). In a canyon west of the collapse, erosion has cut into the top of the Coconino Sandstone through the Kaibab Limestone and the lower shale member of the Moenkopi Formation. These exposures are bleached in places and appear altered. Bleaching extends 1000 feet from the margin of the collapse, but comparatively little structural depression occurs, except close to faults.

The collapse structure trends northeast and surrounds a topographically anomalous hill which has 150 feet of relief. The hill is about 700 feet long by 400 feet wide and has a Moss Back cap of resistant chaotic core; it is encircled by a bleached margin and lies in an elliptical depression 1400 feet by 800 feet. Surrounding the bleached marginal zone of the collapse are three small zones of intense alteration (Fig. 24). Two of these collapse satellites are about 100 feet in diameter and occur in a structural sag. The third outlying feature is a severly bleached area associated with faulting.

Alteration includes bleaching, ferrugination, calcitization, and petroliferous staining. Most of the bleaching is concentrated within the elliptical collapse outline, whereas the calcification and the ferrugination are restricted to the vicinity of shear zones and minor faults in the collapse area.

The structural influence of the collapse is broader than the bleaching effect. The total spread of deformation forms a basin about 1600 feet long and 900 feet wide. The basin slopes are somewhat asymmetrical; dips of 20°–30° are common on the northeast flank, and some are as high as 45°. On the southeastern flank, dips are commonly 15°–20° and rarely attain 25°. Since the regional dip is to the southeast, the basin is more symmetrical than it appears to be.

The subsurface extent of the depression is concealed by the core of displaced Moss Back Sandstone. The minimum vertical sag of beds *in situ* is about 60 feet, but on the northwest flank the apparent sag approaches 100 feet in the Moenkopi beds between the margin of structural deformation and the contact with the core.

The satellite collapses exhibit similar structural depressions on a reduced scale. The larger of two satellites lies along the axis of the main basin on the southwest end. The surrounding Sinbad beds dip toward the center of the satellite in a basin about 200 feet long, which is aligned with the main collapse and has a sag of 15–20 feet.

The smaller satellite, which has a similar structural sag, lies on the southeast flank of the major structure. A highly bleached area of Moss Back Sandstone forms a displaced core. This smaller structure is about 100 feet in diameter, and the sag of the *in situ* beds is probably about 10–20 feet.

Two lithologies of Moss Back Sandstone are represented in the chaotic core of the main collapse: brown sandstone, which probably represents the upper Moss Back, and gray to white sandstone which may be lower. Generally the brown sandstone occurs as individual large blocks, or groups of blocks along the contact of the core, with the intensely bleached Moenkopi sediments; the gray sandstone occurs in the central core. Blocks within the core are brecciated along borders. The brown sandstone contains considerable added iron and dead oil, whereas intense beaching is present in the gray

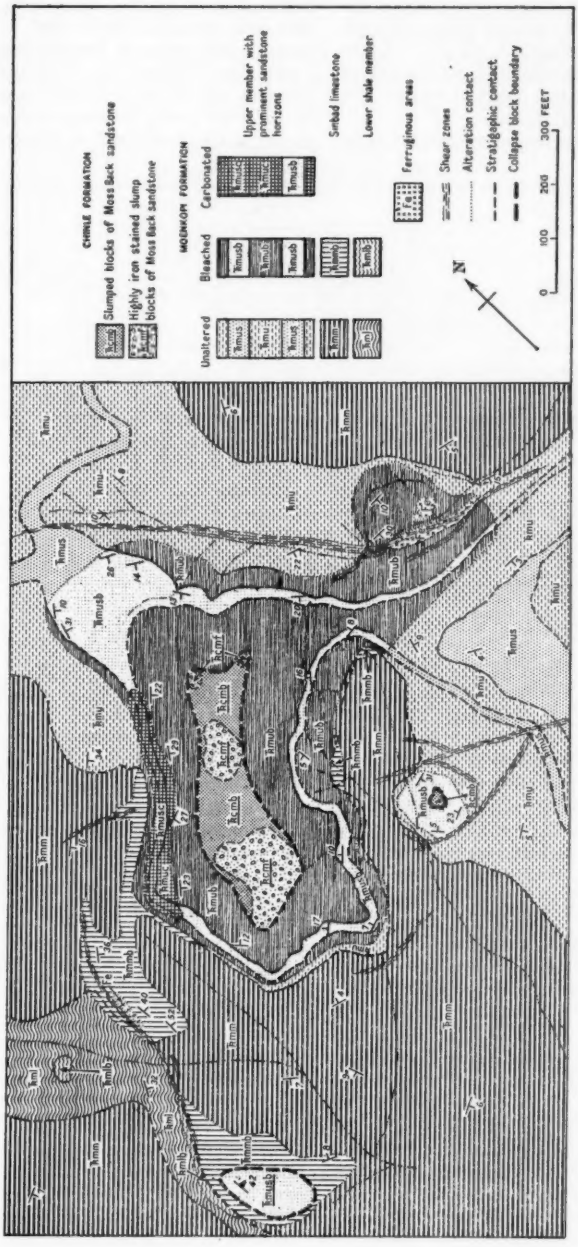

FIGURE 24.—ALTERATION MAP OF COLLAPSE NO. 4

sandstone of the core. The contact between the core and the intensely altered Moenkopi sediments is poorly delineated, since in many places the intense bleaching of the Moenkopi Formation destroys the bedding, so that the sediments are badly fractured.

The collapse core is ovate and has the same trend as the basin. It is 100–150 feet wide, and its long axis is 400 feet. Figure 25 shows the relationship of the Moss Back core to the Moenkopi strata. The upper Moenkopi in a measured section near Temple Mountain is 450 feet thick. The upper Meonkopi sediments, which surround the core in places, extend 80 feet above the Sinbad Limestone Member. A displacement of about 380 feet would be required to drop the core of the Moss Back Sandstone into the present position.

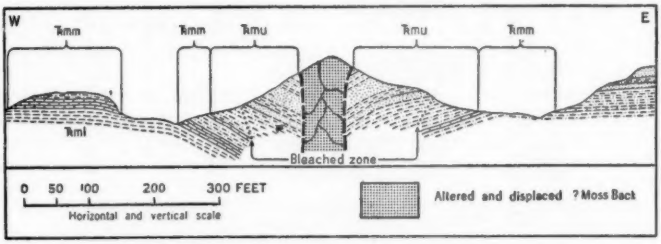

FIGURE 25.—CROSS SECTION OF COLLAPSE NO. 4
(*See* Fig. 24.)

Shear zones are common in the bleached area surrounding the core. Intensely fractured belts are 1–50 feet wide and in places are accompanied by alteration. They generally show little or no displacement, although the rocks are badly fractured and deformed. Where the zones lie in shales some appear as belts of vertical beds a few feet wide and have no displacement from one side of the zone to the other. The shear zones follow the regional joint pattern and tend to form arcuate areas concentric with the collapse structure. The prominent zone on the northeast flank of the collapse (Fig. 24) follows the regional joint pattern, whereas the zones on the northwest and some of those on the southeast form concentric belts around the main collapse or the satellites. Alteration is generally concentrated along these zones, and ferrugination is localized by the shear zones. Small but high concentrations of limonite typify the ferruginous remains and probably come from a pyrite source, as confirmed by the pyrite nodules near these zones. This is particularly evident along the main zone on the northeast flank of the structure. A large mass of intense bleaching protrudes from the general ovate pattern along this zone and on the west flank of the collapse. These shear zones, although less intense, are similar to the shear zones on Temple Mountain and probably represent zones of weakness which ruptured during the formation of the collapse.

Alteration at collapse no. 4 appears similar to the argillic alteration at Temple Mountain in that carbonate and ferric iron have been removed. Bleaching forms halos surrounding the main core and satellite collapses and is also present along most of the shear zones.

Cementation by calcite is localized along the western flank of the structure and is essentially confined to one sandstone bed of the upper Moenkopi Formation, although it spreads for a few feet into the underlying and overlying shales and siltstones. This may represent a concentration of calcite which has been removed from the bleached beds.

Petroliferous material is prominent in the chaotic core and in the Moenkopi sediments. Asphaltic pellets occur in the Moss Back Sandstone, whereas flowing tar seeps occur (Pl. 3, fig. 2) in the Sinbad Limestone Member. Much of the Moss Back contains oxidized petroliferous material. Oil stains occur in the bleached Moenkopi sediments surrounding the core.

Pyrite is common in the Sinbad in massive nodules up to 2 inches in diameter near the tar seeps east of the core. Small amounts of pyrite are distributed through the section from the Kaibab through the Sinbad Limestone in the canyon to the west. Pyrite is fairly common over the San Rafael Swell and is not restricted to the collapse phenomenon; however, where it has been concentrated in sufficient amounts to form gossanlike masses of limonite along the

shear zones, a connection with the collapse is apparent.

Although collapse no. 4 has been considered a prospect, no uranium mineralization has been reported. A radiometric survey with scintillation equipment showed no appreciable radioactivity either along shear zones, within the core, or within the limits of the structural deformation.

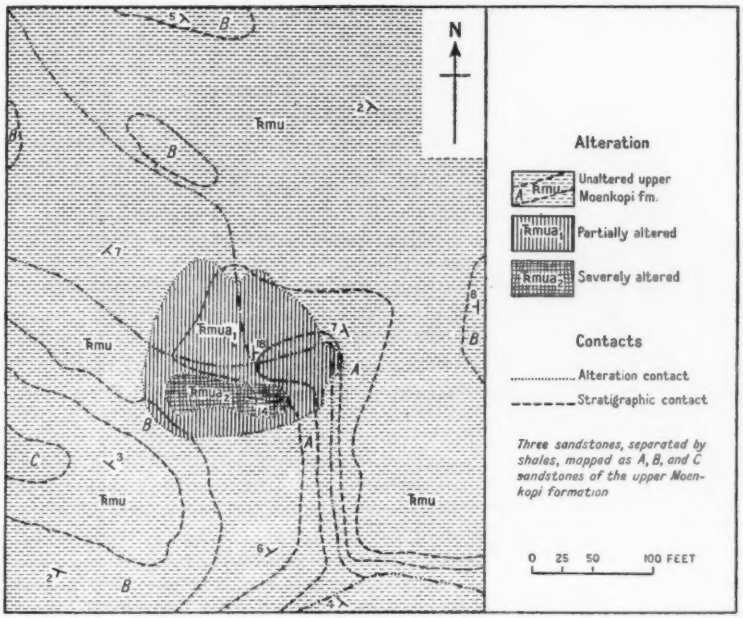

FIGURE 26.—ALTERATION MAP OF COLLAPSE NO. 5

COLLAPSE NO. 5: Collapse no. 5 illustrates a phase of the collapse phenomenon in which the core is not displaced. It is a small feature located half a mile west of the flopover at Temple Mountain. It is circular and about 150 feet in diameter. The country rock is the upper member of the Moenkopi Formation, probably F and G beds (Fig. 19). They are brick-red to tan-brown sandstones, siltstones, and shales, but along the margins of the collapse the beds change to pale tan or almost buff. In the center of the collapse the altered beds are white to light gray on weathered surfaces and dark gray where freshly broken. The strongly bleached zone is about 80 by 30 feet and elongated eastward (Fig. 26).

The regional dip is less than 5° southeast, although a general dip is difficult to determine because of minor flexures. Near the bleached zones the beds abruptly change in attitude and in all cases dip toward the center. Dips surrounding the center of the collapse in the area of marginal bleaching are about 3°–8°; within and adjacent to the intensely bleached zone the dips are up to 18°. Although there is no distinct core, the rocks composing the intensely bleached center are jumbled and show abnormal dips. The rock of the core is a fine- to medium-grained sandstone and is typical of the Moenkopi; hence the presence or absence of a displaced core is not easily established from lithology. The rocks may represent the same sandstone bed which crops out along a canyon near by (Fig. 26), or they could be blocks of Moenkopi Sandstone dropped from higher in the section.

As in collapse no. 7, this collapse does not show the typical displaced core; however, the alteration pattern and structural basin conform. A reconnaissance with scintillation equipment indicated no abnormal radioactivity.

COLLAPSE NO. 6: Collapse no. 6 lies about 4 miles north of Flattop Mesa (Figs. 3, 4), close to the anticlinal crest of the swell. It exhibits

alteration of the country rock, a displaced core of Moss Back sediments, and a basinlike structure. According to the photogeologic map of the United States Geological Survey (1951), the tion noted at collapse no. 4. Much of the highly altered rock where freshly broken is extremely dark or almost black, because of petroliferous material.

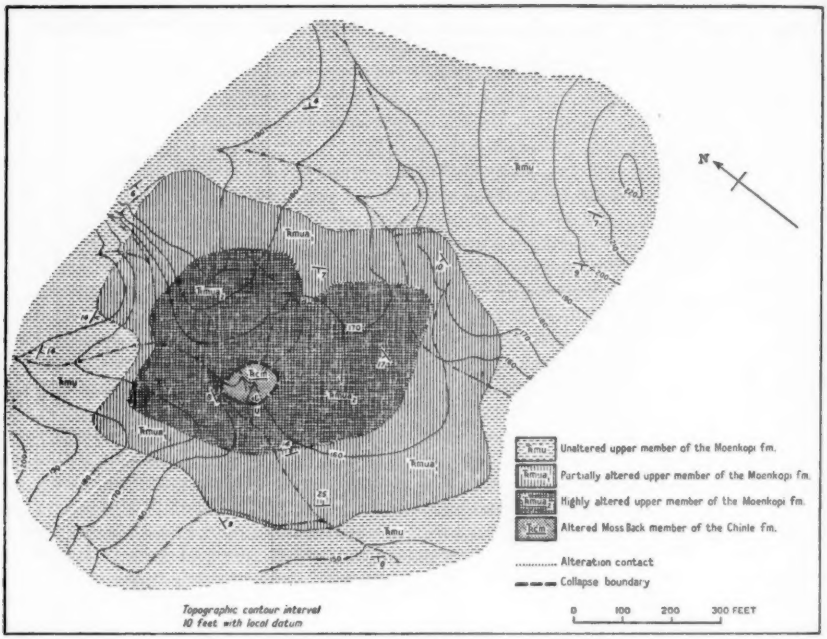
N
Unaltered upper member of the Moenkopi fm.
Partially altered upper member of the Moenkopi fm.
Highly altered upper member of the Moenkopi fm.
Altered Moss Back member of the Chinle fm.
Alteration contact
Collapse boundary
Topographic contour interval
10 feet with local datum
0 100 200 300 FEET

FIGURE 27.—ALTERATION MAP OF COLLAPSE NO. 6

axis of the San Rafael anticline lies less than half a mile to the west.

The altered area measures about 900 feet by 700 feet and is elongated north-south. The displaced chaotic core consists of three small outcrops which form a circle slightly more than 100 feet in diameter (Fig. 27).

The affected rocks include the lower strata of the upper member of the Moenkopi Formation and the Sinbad Limestone. In the center of the collapse the rock is bleached nearly white in a circular zone about 500 feet in diameter; away from the center the bleaching becomes less intense. A light-tan-gray phase forms an alteration rim 30–200 feet wide around the intense white zone and into unaltered tan to light-brown Moenkopi. In local areas enrichment in iron is represented by accumulations of hydrous iron oxides. Local areas of carbonate enrichment occur in veins and as cement. These appear to be less intense and much smaller examples of the ferrugination and carbonatiza-

Structurally the collapse is similar to the other collapse features. A general dip of about 4° to the northeast changes abruptly at the collapse to form a basin with the chaotic core in the center. The dips encircling the collapse average about 10°; on the southern flank they attain 25°. The dip in the direction of the regional dip is steeper.

The core of the collapse consists of blocks of Moss Back Sandstone, as indicated by the chert conglomerate and coarse, dirty sandstone. Evidence of wood trash was also observed. The Moss Back core has been displaced as much as 300–400 feet, since these blocks lie against the lowest units of the upper Moenkopi Formation.

A radiometric survey with scintillation equipment failed to reveal significant surface radioactivity.

On the basis of structural position, the proximity of the collapse to the crest of the

anticline appears to preclude subsidence due to artesian waters.

COLLAPSE NO. 7: Collapse no. 7 is located about half a mile southeast of collapse no. 4 (Figs. 3, 4). It is small and does not possess a chaotic core of displaced Moss Back. The basin structure with the accompanying bleaching identified the feature. The bleached area forms an irregular circle less than 200 feet in diameter and is slightly elongated north-south.

The beds affected by the collapse are the lower strata of the upper member of the Moenkopi Formation. The exposed rocks are stratigraphically 100 feet higher than the upper Moenkopi rocks at collapse no. 4. As in the other collapses, two intensities of alteration were noted. The central or most intensely altered rock is white to light gray, but fresh surfaces are dark gray. This zone (Fig. 28) is about 150 feet long and ranges in width from 80 feet at the southern end to nearly 140 feet at the northern end. The outer zone, in contrast to the normal tan to brick red of the unaltered Moenkopi beds, shows a slightly less intense coloration and forms a band 10–70 feet wide around the central portion. Two small areas of intense cementation of calcite occur within this outer zone. Both are in the southwest corner and measure 15 by 20 feet and 45 by 20 feet. As in the other collapses the central portion indicates some local enrichment in iron.

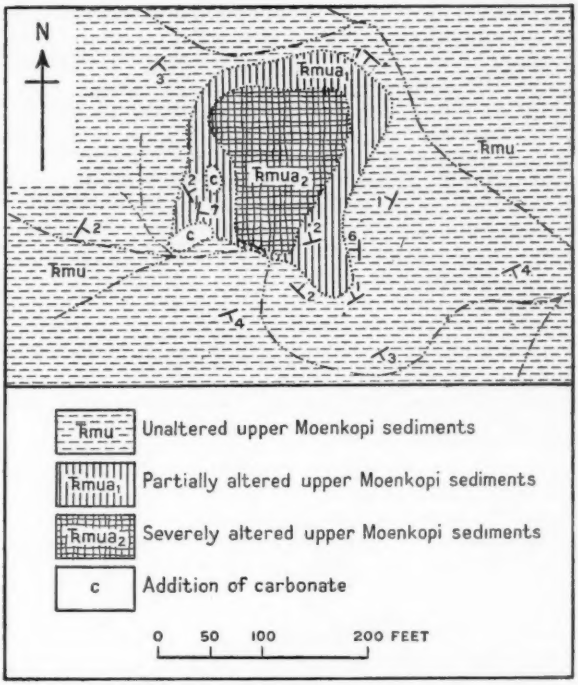

FIGURE 28.—ALTERATION MAP OF COLLAPSE NO. 7

COLLAPSE NO. 8: Collapse no. 8 (Fig. 29) lies on the west side of the San Rafael Swell a few hundred yards from the Wingate escarpment. It is a long narrow feature somewhat less than 1000 feet long and about 200 feet wide. On the north is a long narrow synclinal basin in bleached Moenkopi strata. At the south end is a pyramidal mass of coarse sandstone and conglomerate foreign to the Meonkopi strata of the region.

The source of the conglomeratic strata is not certain. One possible source is the Moss Back Formation which is exposed on the near-by escarpment below the Chinle and Wingate. Another is the Kayenta above the Wingate. The pyramidal block is bleached and stands away from the escarpment; lithologically dis-

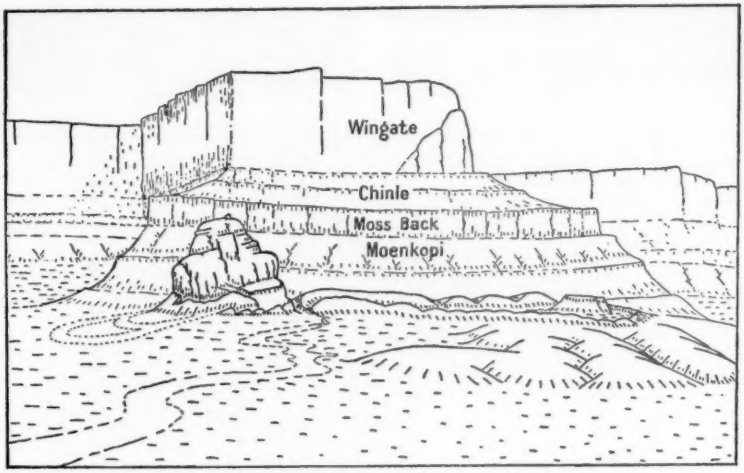

FIGURE 29.—SKETCH OF COLLAPSE NO. 8

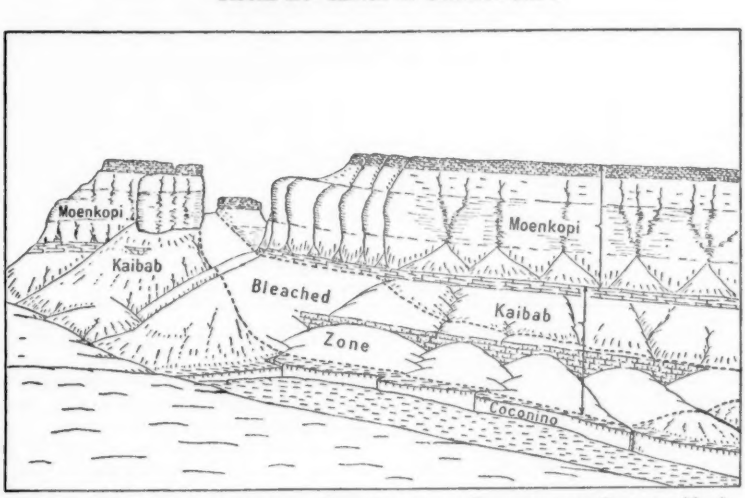

FIGURE 30.—SKETCH OF BLEACHED ZONE EXTENDING OUTWARD FROM COLLAPSE NO. 9
The canyon wall measures about 150 feet vertically.

similar sandstone and shale of the Moenkopi lie between.

This collapse was discovered in an air reconnaissance after the main field work had been completed and was not studied in detail. It does not appear to be connected with any recognizable faulting. Near by, however, on the back slope of the west rim of the swell is an anticlinal flexure shown on the U. S. Geological Survey photogeologic map. The collapse lies along an east-west belt which appears to include the collapse features thus far recognized. It exhibits the structural depression, erratic blocks, absence of faulting, and bleaching characteristic of the other collapse features.

COLLAPSE NO. 9: The area designated as collapse no. 9 lies toward the center of the San Rafael Swell in an intermediate position between collapses 1 and 8. Strata exposed at the surface consist of the lower part of the

Moenkopi, the Kaibab, and the uppermost Coconino.

This collapse has not been studied in detail, but a number of significant features were recognized. A large amount of bleaching occurs (Fig. 30) on an escarpment along a canyon wall. The bleaching lies above the Coconino Sandstone and cuts diagonally across the Kaibab and Moenkopi.

In places the Kaibab Limestone has been largely decomposed by the removal of carbonate and the concentration of residual quartz grains. Bleaching was followed by a large amount of secondary action, and gypsum and calcite geodes formed. Bleaching extends outward from faults, as shown in Figure 31.

This collapse area appears to signify alteration to be expected at the stratigraphic level of the Kaibab, where the resistant strata of the Coconino are too close to permit structural basins which characterize collapse features in higher strata.

LITTLE JOE AREA (SULFUR SPRING CANYON, COLLAPSE NO. 10): The Little Joe area lies in the vicinity of Green Vein Mesa and the Lucky Strike mine along the western flank of the San Rafael Swell, approximately 14 miles northwest of Temple Mountain (Fig. 32). It is almost directly across the anticline from Temple Mountain. The warm spring, urano-organic ore, and structural depression which occur here are associated with a fault which cuts the west limb of the San Rafael Swell.

The Little Joe fault strikes nearly east for about 7½ miles from Green Vein Mesa on the east across the escarpment of the swell on the west. The fault is high-angle normal throughout most of its length; the north side is higher than the south side. Displacement is as much as 300 feet locally east of the Little Joe area. At the Little Joe prospect the main fault is distributed into at least three smaller faults which constitute a fault zone nearly 500 feet wide. Two are high-angle reverse faults, and the third is a low-angle normal fault. Displacement is about 50 feet, and the north side is higher than the south side.

Near the eastern intersection of the smaller faults with the main fault a small outcrop of brecciated Wingate contains urano-organic ore. To the west the smaller faults rejoin at a point near a small warm sulfur spring. The fault zone extends across the northern edge of a narrow structural trough which strikes eastward into the Little Joe area. Where the trough is corrected for a regional dip of 3° to the northwest the fault zone lies along the north flank of a broad depression centering in Wingate Sandstone approximately 800 feet to the south (Fig. 32). This depression may represent the surface expression of a collapse at depth.

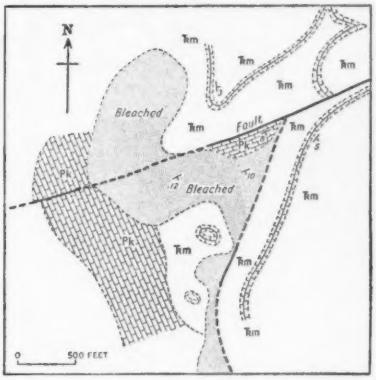

FIGURE 31.—ALTERATION MAP OF COLLAPSE NO. 9

Along the fault a ridge underlain by upper Chinle and lower Wingate sediments shows structural displacement. Valleys to the east and west exhibit lower Chinle and Moss Back on the north, which abut against upper Chinle on the south.

Along the faults red Chinle siltstones and mudstones are altered to green, and tan Chinle and Wingate sandstones are altered white. Locally, along bedding planes and permeable zones adjacent to the faults, friable lower Wingate Sandstone is impregnated with "dead oil". Within the fault zone basal Wingate Sandstone is highly indurated in places. Small faults along the western edge of the fault zone in some places contain streaks and patches of nonradioactive asphaltic material in calcite veinlets.

At the spring, the Moss Back Sandstone locally yields radioactivity of 10 times background. One strip sample yielded 0.02 per cent U_3O_8, 0.06 per cent V_2O_5, 19.9 per cent $CaCO_3$, and 0.09 per cent Cu. The temperature of the spring is about 25°C. throughout the year. Hydrogen sulfide gas is noticeable some distance from the locality. Muds and silts in the spring waters are black and may contain hydrocarbons.

The small patch of fractured and distorted lower Wingate Sandstone near the east side of

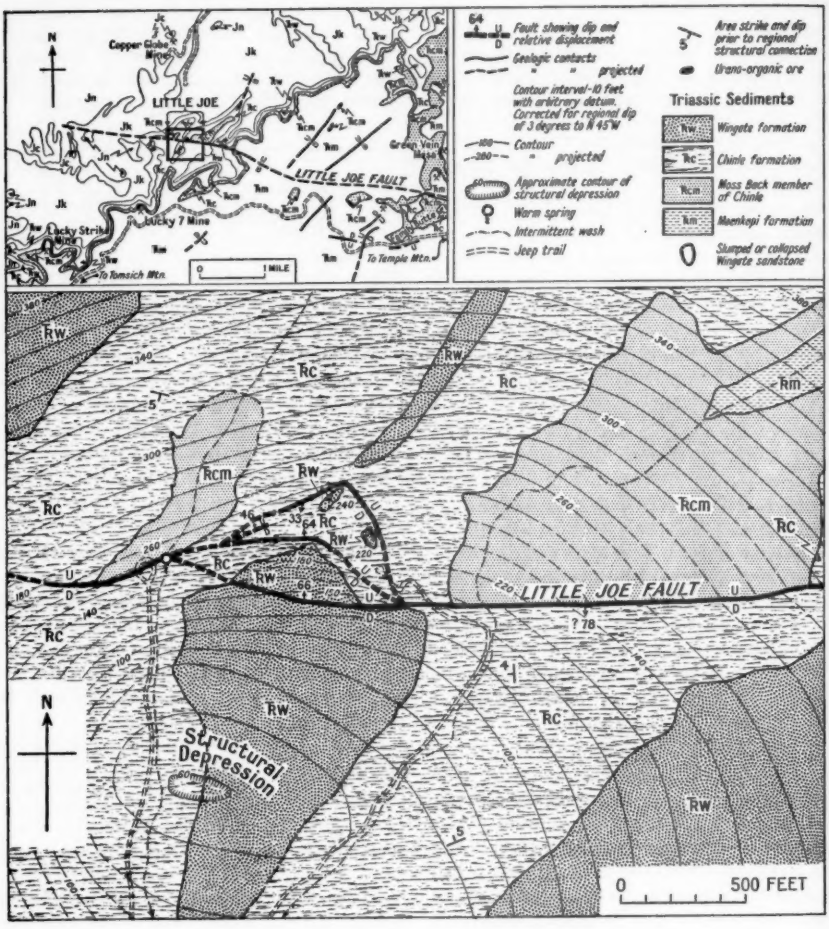

FIGURE 32.—OUTLINE OF COLLAPSE NO. 10

the fault zone is similar to material that constitutes the collapsed core in other collapse areas. Its position in Chinle sediments could be attributed to either slump or collapse. Urano-organic ore is present as pisolites, pods, and streaks along the fractures, bedding planes, and lithologic contacts. Secondary uranium and copper stains are present on ore surfaces. Two strip samples show 0.86–4.71 per cent U_3O_8, less than 0.07 per cent V_2O_5, 2.8–4.3 per cent $CaCO_3$, and 0.31–0.98 per cent Cu. The amount of copper is unusual for the swell. It may signify a local area high in copper in the vicinity of the abandoned Copper Globe mine near by.

RELATIONSHIP OF THE COLLAPSE FEATURES TO MINERALIZATION

At Temple Mountain both minable concentrations of urano-organic ore and unusual collapse structures appear to be related. The minerals from the Temple Mountain area are listed in Table 1. The list includes minerals identified in previous investigations and a summary of the mineralogical data applicable to the current study.

The original sedimentary detritus includes predominately quartz and varying amounts of orthoclase, microcline, plagioclase feldspar (chiefly oligoclase and andesine), chert, chalce-

TABLE 1.—MINERALS OF THE TEMPLE MOUNTAIN URANIUM AREA*

Metallic Mineralization		Alteration	
Hypogene Mineralization			
Urano-organic material		Kaolinite	$Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$
Uraninite	UO_2	Illite	
Pyrite	FeS_2	Chrome-bearing mica clay	
Marcasite	FeS_2	Calcite	$CaCO_3$
Sphalerite	ZnS	Dolomite	$CaMg(CO_3)_2$
Native arsenic	As	Siderite	$FeCO_3$
(?) Arsenopyrite	FeAsS	Hematite	Fe_2O_3
(?) Loellingite	$FeAs_2$	Jarosite	$K_2Fe_6(OH)_{12}(SO_4)_4$
Montroseite	$V_2O_3 \cdot H_2O$	Adularia	$KAlSi_3O_8$
Galena	PbS	Alunite	$K_2Al_6(OH)_{12}(SO_4)_4$
(?) Chalcopyrite	$CuFeS_2$	Opal	$SiO_2 \cdot nH_2O$
(?) Covellite[1]	CuS	Chalcedony	SiO_2
(?) Bornite[1]	Cu_5FeS_4		
Supergene Mineralization			
Metatorbernite[1]		$CaO \cdot 2UO_3 \cdot As_2O_5 \cdot 8H_2O$	
Zeunerite[1]		$CaO \cdot 2UO_3 \cdot As_2O_5 \cdot 8H_2O$	
Garnotite[1]		$K_2O \cdot 2UO_3 \cdot V_2O_5 \cdot 3H_2O$	
Tyuyamunite[1]		$CaO \cdot 2UO_3 \cdot V_2O_5 \cdot 3H_2O$	
Abernathyite[3]		$K_2(UO_2)_2(AsO_4)_2 \cdot 10\text{–}12H_2O$	
Uvanite[4]		$2UO_3 \cdot 3V_2O_5 \cdot 15H_2O$	
Rauvite[1]		$CaO \cdot 2UO_3 \cdot 5V_2O_5 \cdot 16H_2O$	
(?) Uranospinite		$Ca(UO_2)_2(AsO_4)_2 \cdot 8\text{–}10H_2O$	
Corvusite[1]		$CaO \cdot V_2O_4 \cdot 5V_2O_5 \cdot 16H_2O$	
Metahewettite[1]		$CaO \cdot 3V_2O_5 \cdot 11H_2O$	
Pascoite[2]		$2CaO \cdot 3V_2O_5 \cdot 11H_2O$	
Pintadoite[2]		$2CaO \cdot V_2O_5 \cdot 9H_2O$	
Erythrite[1]		$Co_3(AsO_4)_2 \cdot 8H_2O$	
Realgar		AsS	
Orpiment		As_2S_3	
Gypsum		$CaSO_4 \cdot 2H_2O$	
Jarosite		$K_2Fe_6(OH)_{12}(SO_4)_4$	
Alunite		$K_2Al_6(OH)_{12}(SO_4)_4$	
Copiapite[2]		$Fe_4(OH)_2(SO_4)_5$	
Halotrichite[2]		$FeSO_4 \cdot (Al_2SO_4)_3 \cdot 24H_2O$	
Celadonite[2]		Fe, Mg, K silicate	
Brochantite[1]		$CuSO_4 \cdot 3Cu(OH)_2$	
Kaolinite		$Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$	
Illite			
Calcite		$CaCO_3$	
Magnesite		$MgCO_3$	
Goethite		FeO(OH)	
Limonite		$Fe(O, OH) \cdot 2H_2O$	
Sulfur[2]		S	

* If not otherwise indicated, the minerals listed have been identified as part of this study.

[1] Gruner and Gardiner (1952).

[2] Hess (1922).

[3] Thompson *et al.* (1956).

[4] Weeks and Thompson (1954).

dony, and the accessories zircon, apatite, tourmaline, garnet, and sphene. Original and diagenetic interstitial minerals include quartz, kaolinite, illite, sericite, and various iron and hydrous iron oxide minerals. Concentrations of these minerals depend on the stratigraphic horizon and type of lithology. Clay minerals and considerable amounts of carbonate (chiefly calcite) are characteristic of the Kaibab and Moenkopi formations.

Hypogene mineralization.—Considerable alteration within and bordering the collapse features is noted; at Temple Mountain it appears to be associated with the uranium mineralization.

Within the Kaibab and Moenkopi formations the iron oxide minerals and carbonates are most susceptible to change and replacement, and in the upper strata, the Chinle and Wingate formations, the quartz and interstitial clays appear to be the most affected.

Pre-ore alteration.—Alteration minerals associated with the pre-ore phases of alteration include chiefly kaolinite, illite, chrome-bearing micaceous clay (Kerr and Hamilton, 1957), calcite, dolomite, siderite, hematite, jarosite, adularia, and minor amounts of opal. The earliest pre-ore stage is argillization. Here, the carbonates and iron oxide minerals undergo replacement by clay minerals. Argillic alteration is noted within and bordering all the collapse features and appears to become more intense with depth and proximity to the collapse structure. At Temple Mountain, for example, the argillic alteration is most intense within the Meonkopi and Lower Chinle formations bordering the collapse and decreases in intensity outward from the collapse margin. Argillic alteration is consistently present where uranium-bearing hydrocarbon masses have been emplaced. In the field it appears as a severe bleaching of the normally red and brown sediments to a white or light gray.

The second pre-ore stage is well represented only at Temple Mountain and involves the emplacement of carbonate minerals at the expense of detrital quartz. Within the upper Chinle and Wingate formations near the collapse large masses of sandstone have been replaced or partially replaced by dolomite. The introduction of the dolomite follows an orderly sequence which results in an early nodular pattern of interstitial dolomite emplacement gradually intensified to complete replacement of the sandstone. The introduction of calcite follows essentially the same sequence and appears to be earlier than the dolomite.

The final pre-ore alteration stage, again confined to the Temple Mountain area, involves the introduction of iron in the form of siderite and hematite. This results in weathered replacement bodies which closely resemble gossans and occur in the upper Chinle and Wingate formations surrounding the Temple Mountain collapse. The ferrugination is later than the dolomitization and in several places cuts the earlier carbonates.

In some places within ore deposits near the collapse in the Moss Back Sandstone, jarosite is intermixed with the clays and is earlier than the ore. This mixing may also be part of this final pre-ore alteration stage.

The sequence of alteration indicates that solutions, probably acidic rise from below, remove the carbonate material, and deposit clay minerals. On reaching the higher strata, however, the solutions become alkaline and cause the deposition of carbonate and iron minerals at the expense of the quartz.

Metallic mineralization.—Following the climax of the early alteration stages deposition of metallic minerals and the formation of the uranium hydrocarbons occurred. The ore consists of uranium hydrocarbon, similar in physical characteristics to thucolite, and minor amounts of included pitchblende (Kerr and Kelley, 1956). Associated metallic minerals include abundant pyrite, minor marcasite, sphalerite, and native arsenic. Chalcopyrite, covellite, bornite, and possibly arsenopyrite are present in places. Galena becomes more abundant with depth and is most noticeable in drill cores from the upper Coconino and Kaibab within the collapse zone at Temple Mountain. Occurrences of native arsenic and sphalerite (Kerr and Lapham, 1954) intimately associated with the ore have been found only near or within the collapsed sediments. Marcasite has been identified only from cores of drill holes in the collapse.

Various forms of nonuranium-bearing hydrocarbons are present in the collapse areas and include fluid and semifluid tar, dead-oil saturated sandstones, and a hard, brittle nonuranium asphalt tentatively identified as uintaite, a variety of gilsonite found in eastern Utah (Abraham, 1920; Crawford, 1948, p. 115-118). The major occurrences of these hydrocarbons show certain variations from those of the urano-organic ore. Field and laboratory evidence suggests movement of oil from the time of formation of the structure to the present. Oil pervades permeable strata in many places, over a wide vertical range on the swell. Adja-

cent to the collapse at Temple Mountain oil sands have been found in sands in drill holes from the Coconino to the upper Wingate Sandstone. In the district the asphalt (uintaite) has been recognized only at Temple Mountain, although other localities on the swell may exist.

There are two recognizable types of uranium-bearing organic material: one is clear and light gray on polished surfaces and is partially translucent in some thin sections; the second appears turbid and darker gray on polished surfaces and is opaque in thin section. The opaque variety is generally centrally located in masses of ore. In addition to the pitchblende, minute metallic grains of arsenic and pyrite are included within the urano-organic ore which are indicated in X-ray diffraction patterns and electron micrographs. These crystallites cause the turbidity. In spectrographic analyses (Erikson, Myers, and Horr, 1954) the ash of petroliferous material from sands near Temple Mountain show a characteristic suite of elements including arsenic, copper, and molybdenum. Vanadium is relatively low. There appears to be a slight increase of Fe, As, Cu, and Pb and a decrease in V in the ash of the uranium-bearing organic material compared with the nonuranium-bearing material. X-ray fluorescence of the uranium urano-organic ore exclusive of visible pitchblende particles shows that the uranium concentration of analyzed specimens ranges between 3.3 and 8.4 per cent equivalent U_3O_8; iron is moderately abundant, and arsenic, vanadium, chromium, and zinc are generally present in concentrations above trace amounts. Arsenic and vanadium are present in the coalified wood fragments of Temple Mountain, and arsenic occurs in at least trace amounts in all the Temple Mountain organic material tested. Major trace elements are partially reflected by the metallic minerals and suggest restricted localities of introduction rather than regional emplacement.

Field observation and laboratory data suggest the progressive development of masses of ore. Initial weak mineralizing solutions infiltrated the permeable sediments that contain organic matter. The permeation of the interstices between sand grains and the replacement of interstitial cement was accompanied by introduction of the uranium and coagulation of the organic matter present. As the strength of the mineralization increased, the small nodular forms and streaks developed about some nuclei of pyrite or clay, which coalesced to form larger aggregates of massive ore. Replacement of the sedimentary grains and previous interstitial sedimentary and alteration products occurred. Some zoning of the organic matter and associated metallic minerals developed as the intensity, composition, and quantity of the mineralizing solutions varied through time. Coagulation of the colloidally dispersed amounts of precipitating pitchblende and other metallic minerals occurred and was followed by the induration of the hydrocarbon. Polymerization and oxidation were induced by heated solutions. Apparently hardness and uranium content increased with the formation of masses of ore.

Post-ore alteration.—Following the deposition of the uranium and other metallic minerals there was a final phase of probable hypogene alteration in which kaolinite-alunite veins formed in the sediments near the ore. Paragenetic relations indicate that the alunitization is later than the ore; however, it is apparently also earlier than the typical supergene effects.

Supergene Mineralization

Late weathering and downward-percolating ground water yielded a large suite of supergene minerals. Secondary uranium minerals formed at the expense of the urano-organic material and pitchblende are common; they include metatorbernite, zeunerite, carnotite, tyuyamunite, abernathyite, and possibly uranospinite, uvanite, and rauvite. Secondary vanadium minerals include corvusite, metahewettite, pintadoite, and pascoite. Other supergene minerals are jarosite, illite, kaolinite, copiapite, halotrichite, bronchantite, celadonite, erythrite, realgar, orpiment, and calcite. Alunite is present in places as a supergene mineral. Goethite and limonite are abundant as weathering products of pyrite.

Summary of Ore Genesis

The field and laboratory data suggest above normal temperatures for the formation of the ore. Solutions in the area associated with uranium mineralization probably were enriched at least in arsenic, vanadium, and zinc. Conditions that would deposit native arsenic with sphalerite and uranium hydrocarbon indicate solutions at elevated temperatures. Laboratory experiments indicate that the nonuranium fluid tar present at Temple Mountain can pick up uranium from a uranyl sulfate solution under reflux conditions and become partially indurated if heated for a number of days at about 200°–250°C. Also such material as thuchol-

ite is generally considered to be formed under hydrothermal conditions (Ellsworth, 1928; Davidson and Bowie, 1951; Bowie, 1955; Kerr, Rasor, and Hamilton, 1951).

Evidence of hydrothermal activity is also suggested by the alteration of the sediments associated with the collapse features. The resultant suite of alteration minerals is typical of epithermal and mesothermal mineralization (Kerr, 1951; Kerr *et.al.*, 1957; Lovering, 1949; Sales and Meyer, 1948). Iron, magnesium, calcium, and carbonate ions have been removed from the Kaibab, Moenkopi, and lower Chinle strata and emplaced at higher levels in the upper Chinle and Wingate as large masses of dolomite and siderite. At the same time the lower strata underwent extensive development of illite and kaolinite. This activity was followed by the climax of sulfide mineralization with uranium deposition.

A considerable vertical distribution of ore amounting to more than 1500 feet stratigraphically occurs near and within the collapse area at Temple Mountain. Ore occurs particularly along fractures within and surrounding the collapse where earlier alteration had taken place, and the ore is limited to the permeable zones within the Moss Back Sandstone; this occurrence points toward the close association of the collapse feature and the alteration-mineralization sequence. Several factors point toward the intimate relationship between the slumping, alteration, and uranium mineralization: (1) The contemporaneous relationship of the collapse feature and the alteration are shown by the combined brecciation, alteration, and mineralization beneath the collapse blocks, which in some places show an altered matrix and in other places only altered fragments. (2) Intensity of the argillic alteration decreases away from the collapse features. (3) Native arsenic and sphalerite are present only within the collapse or near by. (4) There is a broad stratigraphic range of nonuranium-bearing hydrocarbons throughout the swell, but urano-organic materials occur in local areas. (5) Dolomitization and sideritization are restricted to the vicinity of the collapse zone or local areas. The above suggest that the collapse structure acted as a conduit along which the hydrothermal alteration and mineralizing solutions moved upward. Scattered precipitation occurred along the conduit, but there was considerable lateral migration, particularly in the Moss Back.

Theories of Collapse Origin and Ore Emplacement

Explanation Advanced

The interpretation proposed here postulates rising thermal solutions, probably as an after effect of igneous activity. The elements Ca, Mg, and Fe were leached from Triassic strata. Overlying strata slumped into the caverns created. There ensued argillic alteration, calcification, dolomitization, sideritization, pyritization, metallic mineralization, reaction of uranium-bearing solutions with organic matter, and subsequent erosion. A discussion of this explanation and alternative explanations which have been offered follows.

Landslide

A double landslide was proposed by Gruner (1951) to account for the displaced blocks of Wingate and Moss Back strata of the flopover. In conjunction with the landslide theory, it has even been suggested informally that the petroliferous material acted as a lubricant for the sliding of the large blocks, in much the same fashion as Kugler (1933) attributes the presence of petroleum as a factor in "sedimentary volcanism." Although the landslide explanation was subsequently withdrawn by the author (Gruner *et al.*, 1953), the inadequacy of a double landslide to account for the collapse merits attention.

The drilling program, which has established the vertical drop of Wingate and Moss Back strata to positions of deep burial beneath the present surface, furnishes the most convincing evidence of the absence of surficial landslides at Temple Mountain.

Sink Hole

The sink-hole theory attributes collapse features to limestone-cavern development. Sink holes and poorly developed karst topography occur in areas underlain by Kaibab strata in the northern part of the swell. The phenomena, however, appear to be related to recent drainage and are not comparable to such areas as reported in Missouri (Bain, 1901; Dake, 1923) or South Carolina (Cooke, 1936, p. 80) where rainfall is greater. Cavern slumping with accompanying brecciation in sink-hole development could produce possible collapse features. However, the upward movement of

Ca, Mg, and Fe in great quantity is difficult to reconcile with supergene normal sink-hole formations. It is also improbable that solution and slumping are part of the present cycle of erosion. The example given shows absence of argillic alteration. Sink holes exist on the swell (Pl. 3, fig. 3). However, the consistent occurrence of argillic alteration with collapse features is significant. Migration of solutions, as indicated by calcification, dolomitization, sideritization, pyritization, and metallic mineralization into strata largely above the carbonate beds indicates upward rather than downward movement of solutions.

According to one modification of the sink-hole theory, collapse is attributed to the solution of salt in the underlying Paradox Formation with subsequent slumping. The evidence of wild-cat oil drilling in the vicinity of the San Rafael Swell, however (Wengard and Strickland, 1954), indicates that the salt mass of the Paradox basin lies east of the San Rafael Swell, and the salt strata are missing in this area.

The Temple Mountain drilling program indicates that the collapse breccia is confined to strata above the Coconino Formation. The latter provides a resistant sandstone unit about 700 feet thick beneath the collapse features, apparently broken only by local faults and not involved in the collapse. If salt exists below and has been dissolved to produce caverns, which is unlikely, the effects have not been transmitted through the Coconino Formation.

Mechanical Deposition

The mechanical-deposition theory introduces uranium by erosion, transportation, and mechanical accumulation. It is supported by the presence of nodules of urano-organic ore which in early work on the swell were interpreted as pebbles.

The nodules are a feature of precipitation connected with the migration of solutions. They may occur in porous sands which are generally devoid of gravel accumulations and are in a variety of stratigraphic positions not limited to gravel horizons. In places they represent precipitation along veins. They grade into irregular or elongate forms inconsistent with mechanical transport and deposition.

The derivation of a considerable part of the uranium from chert pebbles seems even less tenable. Only a few chert pebbles have been found in which the possibility of original uranium minerals exists. These probably represent fracturing and later filling long after the pebbles were deposited in their present site. A leached-pebble origin fails to account for veins of urano-organic ore in the overlying Wingate; it is also inconsistent with the occurrence of urano-organic ore throughout an extended vertical distribution both above and below the pebble beds in the Moss Back.

Ash Leach

Waters and Granger (1953) suggested that small amounts of uranium in bentonite yielded uranium ions by leaching due to meteoric water with the result that deposits have accumulated in the Moss Back below. The theory has been applied most extensively in the Wyoming-South Dakota area (Denson *et al.*, 1954), although it is not accepted by all workers even in this area (King, 1955).

At Temple Mountain the ash-leach theory encounters several serious difficulties: (1) Extensive scintillometer traverses fail to indicate significant radioactivity in argillaceous and possibly partly bentonitic areas of the Chinle Formation. (2) Uranium deposits occur stratigraphically above and below the horizon where bentonite is ordinarily found. (3) Argillic alteration borders the collapse features or faults and extends above and below possible bentonite. (4) Large quantities of Ca, Mg, and Fe have been carried upward from strata below the Chinle into overlying Wingate; this indicates an upward rather than a downward surge of solutions. (5) Mineralogical evidence indicates temperatures of formation in excess of those to be expected from downward-percolating waters.

Ground Water

Butler (1920, p. 607–608) and Gruner (1951; 1953) proposed that uranium derived from ground water is responsible for the Temple Mountain uranium deposits. Presumably water would pass over uranium sources in exposed basement areas, pick up traces of uranium ions, migrate great distances, and ultimately precipitate uranium minerals in sites like those near Temple Mountain.

The paleogeography of this situation would require large available areas of basement containing uranium over which water would flow. These areas would have been far removed from

the San Rafael Swell following Laramide deformation when the ores accumulated.

Only partial information exists concerning the configuration of the ground-water profile when ore deposition took place. However, the configuration must have been subject to control by the anticlinal structure of the swell regardless of the depth of sedimentary cover or the amount of denudation. Thus ore deposition due to ground water favors the flanks of the fold and should exhibit reasonably uniform distribution. Exploration has been unsuccessful on the lower slopes of the fold, but ore has been found high on the anticline, as at Flattop Mesa. Also, erratic distribution indicating local centers is a feature of mineralization.

Temperature data involved in the mechanism of formation of the urano-organic ore, the associated metallic mineralization, and the mica-clay polymorphs involved in alteration suggest temperatures above those ordinarily associated with ground water. Data at hand point to ore formation under the influence of heated solutions at temperatures in the range 100°–325°C.

Gruner (1956, p. 507) attributes the removal of limestone masses at depth to CO_2 under considerable pressure, which was present in ground waters at Temple Mountain. He also suggests that high pressures associated with gas and oil structures may be responsible for the collapse structures.

The hydrostatic pressure is inferred to be substantially less than the rock pressure. The latter on the basis of maximum static load was probably at one time not less than 10,000 p.s.i. and not more than 15,000 p.s.i. The corresponding temperature range on the basis of thermal gradient (neglecting chemical action) might have been 100°–150°C.

The solution of limestone would release CO_2 which would be present in solution. The upward migration of Ca, Mg, and Fe appears to be accepted by Gruner (1956), but it is incorrect to assume that the precipitation of U, Cr, As, Zn, and Pb occurred during this early part of the emplacement. Uranium and associated metallic elements and a considerable amount of pyrite were deposited after the carbonate replacement bodies were formed. An abundance of H_2S must have been present in the vegetation of the Moss Back and associated with petroliferous matter. Apparently uranium was deposited in an environment high in H_2S and subject to the reducing action of organic matter subsequent to the action of CO_2.

In the absence of a more detailed explanation it is difficult to attribute the collapse structures to the high pressures associated with gas and oil structures, as suggested by Gruner (1956). The general distribution of the collapse features is better known than it was when his paper was written. They follow a fracture pattern which cuts across the anticlinal structure east-west. If high gas pressure were responsible the collapse features would follow a pattern related to the anticlinal axis. Oil-field blowouts due to gas and resultant collapse structures should be common if the proposed mechanism were generally prevalent.

Breccia Pipe

In places fluids containing gases have forced their way upward through sediments and formed vertical pipelike bodies. The angular fragments produced may be cemented with fine quartz, clay, mica, or metallic minerals including uraninite (or pitchblende). The boundary between the fragmented sedimentary rocks and the undisturbed strata marks the edge of the breccia pipe. The pipes are commonly circular or elliptical in plan and vertical with parallel sides in projection.

Breccia pipes containing copper and silver are common. Breccia pipes containing uranium deposits are known in three places in the northern Arizona-New Mexico region: the Woodrow mine near Laguna, New Mexico (Gruner, 1956), the Orphan mine along the south rim of the Grand Canyon, Arizona, and a recently discovered locality near Cameron, Arizona.

Brecciation of the pipes is largely attributed to the turbulence of gas action. This mechanism apparently applies, among other places, to the pebble dikes at Eureka, Utah (Farmin, 1934), in part to copper deposits at Cananea, Mexico (Perry, 1933; Valentine, 1936), the breccia core in the copper deposit at Santa Rita, New Mexico (Kerr *et al.*, 1950), and Rancagua, Chile (Lindgren and Bastin, 1935, p. 457–472). Gaseous turbulence may account for the broken material and the upward transportation of fragments, but a later phase of hydrothermal solutions is probably closely related to ore deposition.

Small areas of breccia occur in two places at the surface at Temple Mountain. One of these lies about midway on the west slope north of Sugar Loaf, the other occurs on the east slope south of the nave. In each case a large block of Wingate Sandstone rises about 75 feet vertically

as a cap above the breccia. The breccia on the east slope has been penetrated by a small adit from which urano-organic ore has been mined. The breccia on the west slope appears to be barren.

The ore-bearing breccia consists of angular sandstone fragments, a fraction of an inch to a foot thick, which are cemented with fragmental sandstone, some clay, and a little mica. Ore occurs in fragments and matrix. Both occurrences appear to have been subjected to the action of solutions that cemented the breccia after fragmentation.

In a preliminary report on Temple Mountain, the formation of the breccia was attributed to slumping in connection with the collapse mechanism (Kerr *et al.*, 1955). Shoemaker (1956, p. 161) referred to localities at Temple Mountain assumed by the writers to be these occurrences, as breccia pipes. Shoemaker's interpretation would be reasonable as representing gaseous action bordering the collapse features, but vertical exploration would be required to verify it. Where the breccia is best, but inadequately, exposed in the adit on the east slope, it does not appear to extend downward as one would expect in a breccia pipe.

Since the main collapse at Temple Mountain appears to terminate on a platform of Coconino Sandstone, the accumulation of breccia above is attributed to slumping with subsequent cementation by hydrothermal solutions rather than to the gaseous turbulence of a breccia pipe.

Hydrothermal

The theory of hydrothermal uranium emplacement is considered here in connection with the sequence of events involved in the formation of collapse features and uranium deposits as now found. The sequence involved deposition, deformation, alteration, ore emplacement, and erosion, as follows:

(1) Stratigraphic deposition, Paleozoic through Cretaceous
(2) Laramide deformation and uplift of the San Rafael Swell
  (a) Accompanied by two fracture systems
  (b) Dike and sill penetration on the southwest
  (c) East-west fault pattern
  (d) Initial petroleum migration in porous sediments
(3) Hydrothermal alteration and ore emplacement
  (a) Upward penetration of heated solutions along faults and fractures in isolated centers
  (b) Carbonate migration and argillic alteration in strata below the Wingate
  (c) Calcification, dolomitization, and sideritization in Wingate and upper Chinle
  (d) Pyritization and sulfide deposition
  (e) Uranium emplacement in petroliferous strata
4) Late Tertiary and Quaternary erosion

Cross sections of the collapse at Temple Mountain (Fig. 11) are shown in Figures 12, 13, and 14. Comparative size and stratigraphic range of four of the collapse features are shown in Figure 33. The four are referred to a common datum, the top of the Coconino Sandstone.

In the east collapse at Temple Mountain, blocks of Wingate Sandstone hundreds of feet across are tilted at a high angle and incline radially toward a center. The tops of blocks are partly eroded, and the lower portions are imbedded in the core of the collapse.

An abrupt change in attitude occurs from nearly horizontal strata of North Temple Mountain and South Temple Mountain to steeply tilted beds of the collapse. The large change in angle and the massive brittle character of the Wingate Formation should cause extensive faulting. Although radial faults exists in a number of places, particularly toward the centers of the collapse, large changes in attitude commonly occur without noticeable breaks.

A change in attitude as much as 50° within 50–100 feet occurs without faulting, where the release of pressure appears to have been distributed along minor shear zones, fractures, and bedding. Slumping toward collapse centers probably occurred under considerable load and before the development of the present profile of erosion.

The carbonate content of normal Moenkopi strata, limestone members, and the Kaibab Limestone is large enough where removed to provide space adequate to accommodate the slumped blocks of Wingate and Chinle strata. Abundant carbonate veinlets, considerable thicknesses of limestone, and calcareous shale indicate the wide distribution of calcium carbonate subject to solution.

The siliceous Coconino, about 700 feet thick at Temple Mountain, would not yield solution cavities adequate to accommodate slumped blocks. However, faulting which is largely obscured in overlying strata exists in the Coconino.

Where exposures of the Coconino occur, the formation is massive, resistant to folding, and cut in places by eastward-trending faults. It is not known whether faulting has displaced the Coconino beneath the Temple Mountain collapse or whether an accentuated fracture system has developed. In either case displacement has occurred, and channels were formed which were adequate to permit the migration of solutions upward through the overlying carbonate rocks. Cavernous development in the overlying Kaibab and Moenkopi formations is attributed to the action of such solutions.

Dolomitization and sideritization occur in the Wingate Formation and toward the top of the Chinle Formation close to the collapse.

Urano-organic ore has been found in the old tramway adit (Wingate Sandstone), in fractures in the collapse itself (Wingate and Moss Back) and in the Moss Back Formation at the north and south ends of the section.

Color changes accompany the Temple Mountain collapse. Characteristic reds and browns of the Chinle Formation change to light gray, white, and pale green along the border and outward in places. The common

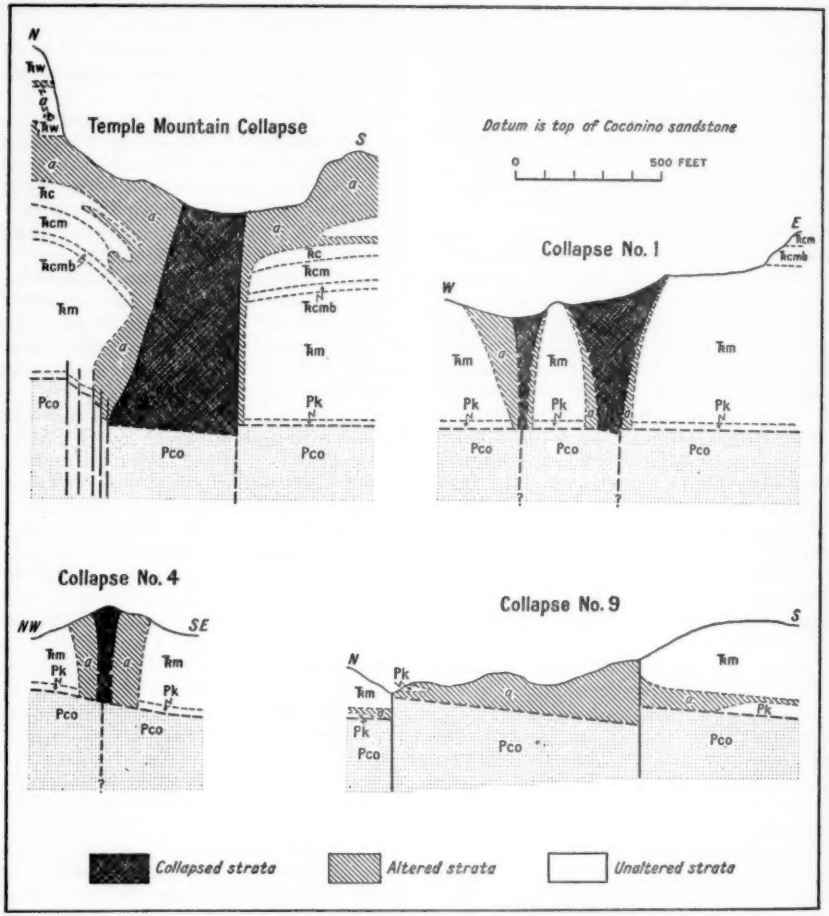

FIGURE 33.—SCHEMATIC CROSS SECTIONS OF COLLAPSE STRUCTURES

hematitic stain has been removed. Clay minerals have been leached and redistributed. Within the large sandstone blocks, thinly distributed dead-oil sand common in the massive normal Wingate strata has been removed. In places concentrations of urano-organic ore occur, and the sandstone appears bleached. Since the action of solutions in terms of Ca, Mg, Fe, and CO_2 ions has been subtractive in the lower horizons and additive above, the movement of solutions was primarily upward.

Evidence is incomplete on the source, temperature, and chemical conditions of the uranium-bearing solutions. Interpretation of the temperatures prevailing at the time of deposition is dependent upon the nature of the resultant mineral products. The principal uraniferous product is a hard, brittle, black urano-organic mass containing specks of pitchblende. Associated with the ore are nodules of pyrite. One nodule studied in detail (Kerr and Lapham, 1954) contained a substantial amount of native arsenic and small amounts of sphalerite and galena as well as pyrite. The temperature of formation in such an association is not established, but temperatures in excess of those due to normal ground water must have prevailed in order to produce hard, brittle asphaltite, native arsenic, specks of pitchblende, sphalerite, and galena.

Pitchblende can be precipitated at 50° C. or probably even at room temperature if given sufficient time. However, precipitation is greatly facilitated by an increase in temperature. At 100° C. it can be formed readily in the laboratory and with even greater facility at a higher temperature (Miller, 1955).

Data on the synthesis of native arsenic are limited, but published experiments favor a temperature on the order of 325° C. Chemical evidence indicates some doubt that it would form at a lower temperature.

Temperatures at which metallic sulfides form cover a considerable range, but they are consistent in the range noted. The weight of evidence favors temperatures in excess of 100° C. and possibly as high as 325° C.

The source of the solutions which deposited the uranium appears deep-seated. Only traces of uranium exist in the dea-oil sands and in viscous tars. The petroliferous materials represent accumulations from a Paleozoic source in permeable strata. Some beds of petroliferous material extend for miles without a single uranium prospect. Tar seeps even in ore zones are nonradioactive. The trace-uranium content of dead-oil sand, where tested, is below the limits which are considered significant. Where dead-oil sands are bleached on the borders of ore bodiès the volume is inadequate to account for the ore on the hypothesis of trace-element accumulation. The most likely source of ore accumulations appears to be hydrothermal solutions produced as a final product of igneous activity.

The nearest igneous rock was found in an oil well at about 4000 feet, 5 miles north of Temple Mountain. The rock was identified as granodiorite. Numerous dikes and sills cut the southwestern and southern slopes of the San Rafael Swell. With one or two exceptions the intrusions are not accompanied by bleaching. The hydrothermal action of the collapse features appears to be independent or later.

A hot spring occurs in the northern part of the swell, and tuffa deposition indicative of former hot-spring action has been observed to the northwest.

At Temple Mountain precipitation appears to have been facilitated by the petroliferous content of the rock penetrated by the heated uranium-bearing solutions. Adequate fractures for vertical migration of solutions exist; migration along strata is an important factor, but there is good reason to believe that the original source was below.

Organic material provided the host for the uranium. A reducing environment prevailed. Precipitation of pyrite at the same time suggests neutral conditions during part of the cycle.

Items indicative of hydrothermal deposition are summarized as follows:

(1) Alteration halo connected with collapse features
(2) Structural pattern of collapse features
(3) Distribution of clay-mineral changes, including mica polymorphs
(4) Hot-spring activity at north end of swell
(5) Dikes and sills at south end of swell which show near-by igneous activity
(6) Polymerized urano-organic material
(7) Upward transport of Mg-Fe and formation of dolomite masses.
(8) Upward Fe transport and formation of siderite bodies
(9) Mineralization yielding chroma mica clay, uraninite, pyrite, native arsenic, sphalerite, and galena

Rising thermal solutions carrying uranium, which encountered petroliferous material and fossil vegetation and precipitated ore, appear to provide the most satisfactory mechanism of

emplacement at Temple Mountain. The source of the uranium ions could be either magmatic fluids or deeply buried uranium-bearing black shale.

A decision concerning the source involves a better understanding of the black-shale distribution and better knowledge of the uranium metallogenic province that embraces the Colorado Plateau. Since underlying black shale is limited in distribution regional considerations appear to favor at least a considerable contribution related to magmatic sources.

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PLEISTOCENE BEDS NEAR THE EDGE OF THE CONTINENTAL SHELF, SOUTHEASTERN LOUISIANA

BY W. H. AKERS AND A. J. J. HOLCK

ABSTRACT

A complete record of Pleistocene deposition is observed in cores from a well at the tip of the Mississippi River delta near the edge of the continental shelf. Marine Pleistocene beds under the tidelands of Louisiana are correlated with the alluvial formations at the surface and in shallow wells farther inland.

A marine transgressive wedge correlatable with the Williana Formation contains Foraminifera which suggest water as deep as 600 feet. A reef-type assemblage in the base of these beds indicates a climate warmer than at present in the northern Gulf of Mexico. Marine sediments correlatable with the Bentley Formation yielded fossils suggestive of depths of deposition as great as 400 feet. Another reef-type deposit is present in the lower part of these beds. Several depositional cycles of lesser magnitude are apparent in the upper part of the Pleistocene section. The last major transgression correlates with the Prairie Formation, and evidence for water as deep as 400 feet over the present tidelands is shown. Still another reef-type assemblage is seen in the lower part of this wedge.

The sandy section below the Williana beds does not present a means of recognizing the base of the Pleistocene, but it may be logically assumed on the basis of the thickness of overlying cyclic deposition to be near 4200 feet. This means that some of the younger beds of offshore Louisiana probably constitute the thickest Pleistocene deposits in the world, if the ages assigned to the alluvial terraces of the lower Mississippi valley are correct.

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INTRODUCTION AND ACKNOWLEDGMENTS

The purpose of this study has been to investigate late Cenozoic sediments as close as possible to the edge of the continental shelf and to determine at that position not only the total thickness of the Pleistocene but also the boundaries, thickness, and depositional environment of its component formations. Such a study was suggested by Trowbridge (1954, p. 805). He summarized the ideas of H. N. Fisk, R. J. Russell, F. P. Shepard, H. C. Stetson, R. S. Dietz, H. W. Menard, and others concerning the alternate exposure and submergence of the continental shelf during the Pleistocene and re-emphasized that the most nearly complete record of Pleistocene history in this country should be found in the sediments of the Gulf Coastal area and the offshore continental shelf. He restated previous ideas that

shore lines during the glacial ages would have been in the vicinity of or beyond the outer edge of the continental shelf, provided only glaciation and deglaciation and the resulting fall and rise of sea level took place. During such times beaches, lagoons, and other littoral features would have existed farther offshore than the present shore line. Trowbridge further reasoned that sea level would have stood farther inland than at present during the long interglacial ages, and that not only the present shelf area but some of the inshore areas would have received sediments and supported a marine environment.

A well drilled near the edge of the continental shelf should encounter marine beds of Aftonian, Yarmouth, Sangamon, Brady, and Thermal Maximum age. These would be separated by nonmarine formations if surfaces were so low and flat during the glacial ages that they received deposits or by disconformities if they were emergent. It was with sampling of these beds in mind that Trowbridge (1954, p. 810) expressed the following:

> "It is hoped that in the near future complete cores from the present surface of deposition to the base of the Pleistocene will be available from drillings on land or in the tidelands of the Gulf Coast. The interpretation of such cores might lead to a more nearly complete history of the Pleistocene that will harmonize with and confirm the history as determined from glacial deposits and interglacial weathering products in the glaciated areas."

The authors accept H. N. Fisk's correlations of the alluvial formations of the Lower Mississippi Valley with the standard glacial and interglacial sequence. By means of subsurface correlations and a faunal study of a well near the edge of the continental shelf a correlation between marine beds and the Pleistocene alluvial section is suggested.

The California Company supplied cores, electrical logs, and laboratory and drafting assistance, as well as all other financial support, for this investigation.

Marine Pleistocene

The subsurface marine sediments of the coastal Louisiana Pleistocene have received meager attention in the literature, although the alluvial sequence has been much described and discussed. The occurrence of marine Pleistocene faunas at shallow depths in a few wells and from several localities was noted by Richards (1939, p. 304–307) and by Anderson and Murray (1953, p. 845). The base of the "Upper Marine Beds", which occurs from depths of 1500 feet in St. James Parish, Louisiana, to more than a mile in offshore Louisiana wells, was contoured by Lowman (1949, p. 1990). He believed that this horizon represented the approximate base of the Pleistocene.

It has not been the practice of operators to catch ditch samples from the upper 5–6 thousand feet of section in southern and offshore Louisiana because of the rapid drilling rate at those depths. Those paleontologists, who have examined cuttings from the few wells in southern Plaquemines, Jefferson, and Lafourche Parishes, and the bordering offshore areas where such shallow intervals were not disregarded, are familiar with rich assemblages of tropical marine fossils representing almost every phylum of the animal kingdom. It was of these sediments that Trowbridge (1954, p. 806) asked "whether the Foraminifera . . . occur in either alternating cold- and warm-water zones or alternating shallow- and deep-water zones."

The objectives of the present writers have been twofold. They have sought an answer to the above questions by comparing the constituents of closely spaced cored intervals with the constituents of modern sediments to determine by analogy the environment of deposition, and they have correlated the sampled intervals with the updip alluvial sequence, thus relating climatic and sea-level changes to the previously described and dated terrace deposits. The cored well at the marine extremity of the investigated area is The California Company, State Lease 2553, No. 1, South Pass Block 41 (Fig. 1). This well was the closest yet drilled to the edge of the continental shelf when the present investigation was initiated. Before presenting the results of our examination of the 158 sidewall cores from the upper 4500 feet of this well, correlations between the cored interval of this boring and the much discussed alluvial sequence to the northwest will be considered.

Correlations, Marine Sediments with Alluvial Formations

The alluvial Pleistocene formations of the Louisiana Gulf Coast were described and named by Fisk (1940) from oldest to youngest as Williana, Bentley, Montgomery, and Prairie. He recognized a unit of essentially sandy lithology overlain by a clay-silt interval in each terrace deposit. In later work (Fisk, 1944) each sandy unit was termed a substratum, and each

FIGURE 1.—SKETCH MAP OF SOUTHEASTERN LOUISIANA SHOWING LOCATION OF WELLS IN CROSS SECTIONS AND THE 600-FOOT ISOBATH ON THE CONTINENTAL SHELF

clay-silt interval the topstratum. A substratum-topstratum couplet has been considered to be a formation. The youngest three of these formations were determined and traced by Fisk (1952) from the outcrops into the subsurface of the Atchafalaya river region (approximately from *B* to *A*, figs. 1, 3).

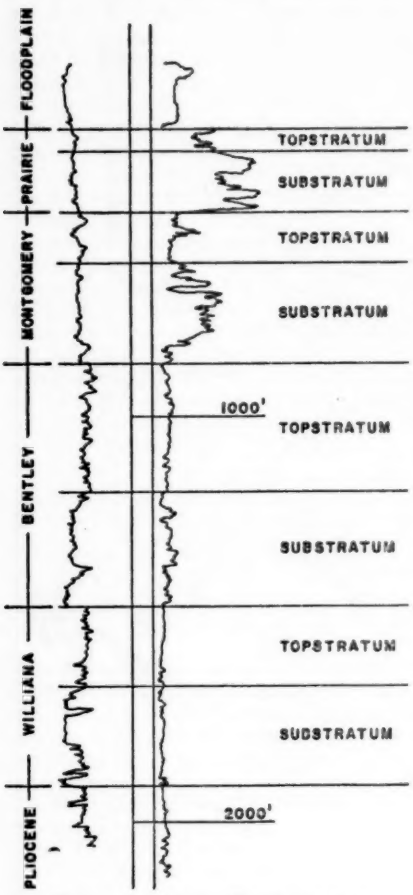

FIGURE 2.—TYPICAL ELECTRICAL LOG OF SUBSURFACE PLEISTOCENE ALLUVIAL DEPOSITS

The writers have examined the subsurface Pleistocene nearer the edge of the present continental shelf (Fig. 1) and through application of the substratum-topstratum cycle have extended the study to greater depth to include what is believed to be the base of the Williana Formation. A typical electrical log of subsurface Pleistocene alluvial deposits is shown in Figure 2. Alluvial deposits are correlated with their marine equivalents by means of the electrical logs of more than 100 wells. Although only a few of these wells are indicated on the accompanying cross sections, they illustrate stratigraphic conditions clearly disclosed by drilling activities of the oil industry in coastal Louisiana.

With the exception of the South Pass Block 41 test, electrical log character was relied on exclusively to determine lithologies. This practice was dictated by the general lack of samples in the upper few thousand feet of section. This lack also necessitated reliance on electrical log character and on the regional development of beds for the determination of environments of deposition (alluvial, transitional, or marine).

In general, alluvial horizons can be determined by the presence of well-developed, thick sands (substratum) separated by a silty, clayey, sandy interval (topstratum) characterized by a highly erratic spontaneous potential curve on the electric log. At the opposite extreme of environment are the marine clays, definitely proven so by foraminiferal evidence in the South Pass test, and characterized, in the area studied, by a relatively smooth spontaneous potential curve. The greatest difficulty in determining environment of deposition obviously is in the zone where nonmarine beds grade to marine. Here in most cases the assignment of a silt and clay unit to any environment is at best a guess, and for simplicity the term transition zone is suggested.

The electrical logs do not offer a means of recognizing unconformities. In areas where beds can be reasonably assigned to an alluvial origin, unconformities have been drawn (heavy lines at the base of each substratum, Pl. 1, 2). The downdip extent of the unconformities is conjectural.

The subsurface basal transgressive interval, called the "Upper Marine Beds", when traced updip by means of electric logs from the tip of the Mississippi River delta to Lapice, about 50 miles west of New Orleans (*C–B*, Pl. 1), is seen to be the marine equivalent of the Williana upper substratum and the Williana topstratum. The immediately overlying sandy regressive section appears to be the seaward continuation of the bulk of the Bentley substratum and includes deposits of pre-Bentley, post-Williana age. Overlying these beds is another transgressive suite correlative with the Bentley upper substratum and the Bentley topstratum. Above this is a regressive interval which corresponds

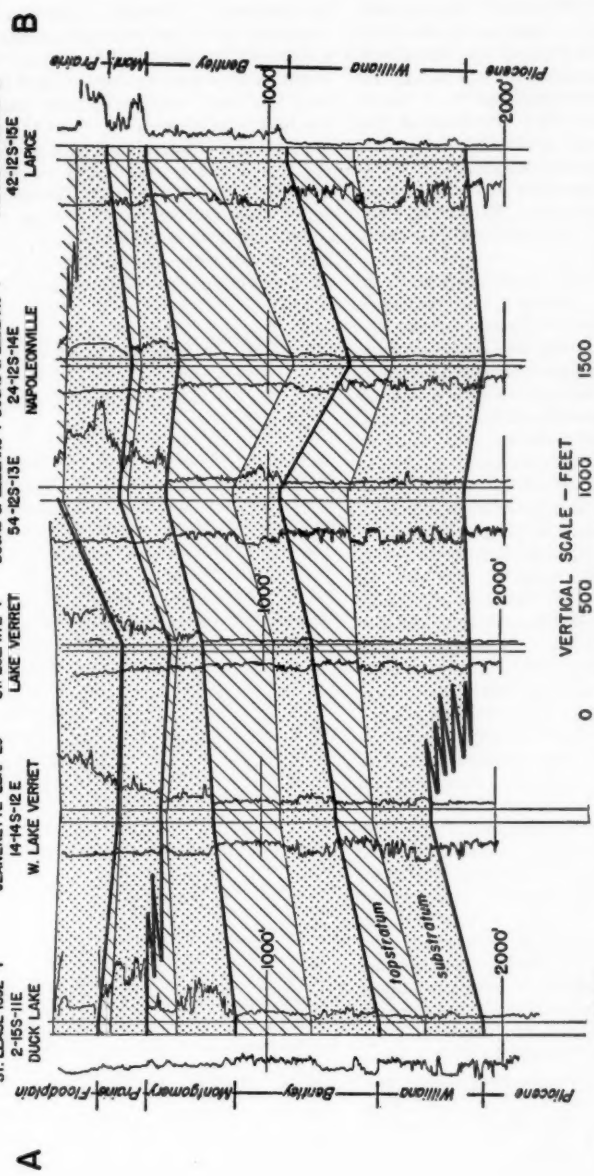

FIGURE 3.—CROSS SECTION A-B, DUCK LAKE FIELD TO LAPICE FIELD, LOUISIANA

with the major part of the Montgomery substratum and probably includes sediments of pre-Montgomery, post-Bentley age. The Montgomery upper substratum and topstratum are equivalent to a marine transgression. A regressive phase correlates with the bulk of the Prairie substratum and probably includes pre-Prairie, post-Montgomery sediments. A transgressive zone correlates with the Prairie upper substratum and topstratum. The Montgomery Formation appears to be a double or possibly a triple sedimentary cycle.

An alternate opinion regarding the age of the shallow beds near the tip of the Mississippi Delta (Southwest Pass) has been presented by Fisk and McFarlan (1955, Fig. 6, section c-c). These authors have mapped the base of the late Quaternary (post-Prairie) deposits at about −800 feet at Southwest Pass. This corresponds to the base of the Prairie Formation as mapped by the present authors (section *B-C*, Pl. 1, well no. 24). That these differences of opinion arise is in large measure because of the difficulty of extending correlations in the subsurface from an alluvial sequence through a transition zone into the marine section definitely developed in the Southwest Pass area. The final solution to this problem offers considerable challenge for future investigation.

Another cross section (*DE*, Fig. 1) extending from a well about 36 miles west of New Orleans southward to Bay Marchand reveals the same equivalencies (Pl. 2), that is, regressions correlate with the bulk of the substratal materials and probably include deposits equivalent in time to the periods of valley cutting in the alluvial section, and transgressions appear to be equivalent to the upper substratum and topstratum of the alluvial formations.

Dating of the marine transgressions and regressions should be determined by the ages of the alluvial terraces. These were dated by Fisk (1952) as Aftonian for the Williana Formation, Yarmouth for the Bentley Formation, Sangamon for the Montgomery Formation, Peorian (later Bradyan, Fisk, and McFarlan, 1955, p. 283) for the Prairie Formation. It has been pointed out by Frye and Leonard (1953) and Trowbridge (1954) that if a rise in sea level were accompanied by alluviation, the terraces are not properly interglacial but postglacial maximum in age. According to this interpretation, the Pleistocene beds of the Gulf Coast may be dated as shown at the right-hand extremity of Plate 1. The recommendation of Frye and Leonard (1953, p. 2585) as followed here proposes the definition of a glacial age as "the time span coincident with the active life of major continental glaciers as expressed in coastal areas by the cycle of fall and rise of sea-level to a position of equilibrium, that is, from the beginning of sea-level fall to the ending of sea-level rise". Interglacial ages would include the time between glacial ages as here defined and would thus comprise the time of essentially stable sea level when neither fall nor rise affects coast lines. Fisk and McFarlan (1955, p. 283) have shown these relationships in the Quaternary stratigraphy of the Louisiana Gulf Coast.

Depositional Environments Inferred from Cores

Sampling Method

Sidewall coring is believed to be the most accurate as well as the most economical method of obtaining uncontaminated samples which can be keyed to depth in the well and position on the electrical log. Ditch samples are notoriously inadequate for ecologic purposes as well as for reference to exact depth in the hole. Incomplete recovery of conventional cores from the unconsolidated Pleistocene sediments would have resulted in only approximate depths for the portions of sediment so obtained. The 158 sidewall cores taken in Pleistocene beds penetrated by The California Company well are plotted on the electrical log of this well (Pl. 1, right-hand extremity of cross section *BC*) and are believed to be accurately positioned with respect to depth and electrical log.

Planktonic Foraminifera

It was hoped that relative temperatures for the cores could be inferred by means of the planktonic Foraminifera, and that alternating warm and cool cycles could be determined for the Pleistocene independently of the water depth factor. Such studies of planktonic Foraminifera have been made with deep-sea cores by Cushman and Henbest (1940), Stubbings (1939), Phleger (1939, 1942, 1947, 1948), and Ericson, Ewing, and Heezen (1952). Although most of the Recent planktonic species were found in the South Pass cores, they offered no satisfactory basis for dividing the Pleistocene according to temperature. Planktonic species which are typical of temperate zones, according to Wiseman and Ovey (1952, p. 65), are found with warm and tropical forms. Either they lived together during at least part of the Pleisto-

cene or the mixing of different water masses resulted in the confluence of their empty shells. The Arctic and Antarctic species, *Globigerina pachyderma* (Ehrenberg) and *Globigerina dutertrei* d'Orbigny, were not found.

Benthonic Foraminifera

Only a few of the benthonic species identified have become extinct before the Recent, and all the foraminiferal assemblages are typical today of some part of the Gulf of Mexico or the Atlantic Ocean. It was soon apparent that some of the cores represented environmental conditions similar to those operative today in the near-shore and inshore waters of the lower Mississippi delta. Others resembled the sediments in several hundred feet of water near the edge of the continental shelf, and a few others contained faunas similar to those now extant in waters of the Florida Keys. Relative changes in sea level were indicated by strong variations in the faunal aspects of the cores, and when these were considered with other constituents of the samples and with the stratigraphic sequence in the area as interpreted by means of electrical logs, marine transgressions, and regressions could be easily recognized. Exact correlations as to water depth on the sole basis of an individual species' bathymetric limits in modern seas is held to be hazardous due to the temperature factor, which is unknown for all the units of the Pleistocene. The environmental curve for the South Pass well (Pl. 1) should be interpreted as a relative measure of depth rather than as absolute determinations, and the depths shown are thought to represent only approximate Pleistocene water depths.

Ecologic Zones

Three major environments or ecologic zones are recognized in the cores. These zones and the gradations between them are present today in the northern Gulf of Mexico. Recent Foraminifera which characterize the various environments have been collected and reported by The California Company (Akers, 1952) and others (Kornfeld, 1931; Post, 1951; Phleger and Parker, 1951; Parker, Phleger, and Peirson, 1953; Parker, 1954; Bandy, 1954; Phleger, 1954). Paleoecologic interpretations of faunal associations by Israelsky (1949) and Lowman (1949, p. 1956) were also applied to the environmental problems of the Pleistocene samples. The three distinctive zones and the criteria by means of which they are inferred are described as follows:

ZONE I, LITTORAL AND SUPRA-LITTORAL: Water bodies are bays, lagoons, or brackish lakes. Coarse quartz sand, plant fibers, and reworked pre-Neogene fossils, mainly *Gümbelinas*, are conspicuous. Clays are also present, and when fossiliferous, they are characterized by *Streblus beccarii* (Linné) variants, *Elphidium* spp., and *Eponidella gardenislandensis* Akers. Other benthonic species are absent or rare, and planktonic Foraminifera are absent.

ZONE II, INNER NERITIC: This environment is the shoreward portion of the continental shelf. The lithology may be clay or fine to medium sand. Echinoid fragments are found with a foraminiferal fauna consisting mainly of *Epistominella vitrea* Parker, *Buliminella* cf. *B. bassendorfensis* Cushman and Parker, *Buliminella elegantissima* (d'Orbigny), and *Bolivina lowmani* Phleger and Parker. Planktonics and benthonic Foraminifera from zones I and III are often present, but they are normally infrequent.

ZONE III, OUTER NERITIC: This environment is the outer continental shelf. Lithologies are dominantly clay and silt, although bodies of fine sand may also occur here. Planktonic Foraminifera are often abundant, and their shells may constitute the dominant part of the assemblage. The benthonic component consists of associations in which the following are dominant or at least conspicuous: *Uvigerina* spp., *Chilostomella* spp., and *Cassidulina* spp. Certain species apparently confined to deep, cool water, such as *Cibicides corpulentus* Phleger and Parker and *Cyclammina cancellata* Brady, are often present.

It is possible to subdivide these zones by means of inferences as to depth of water, but for reasons stated above it is thought that inaccuracies might result by so doing. The zones described above are gradational in the present Gulf of Mexico and also in the Pleistocene section, as they should be in the absence of faulting and major unconformities. The zones as herein defined are indicated opposite the electrical log of the South Pass well (Pl. 1).

Tropical Conditions

Cores at 578, 586, 592, 610, 2418, 3560, 3574-3590, 3668, and 3675 feet contained foraminif, eral assemblages which in modern sediments are often characteristic of fore-reef areas. Approximately 30 per cent of the total number of tests in some of these cores were of tropical planktonic Foraminifera. The last five contained *Globorotalia menardii* (d'Orbigny) var. *miocenica* Palmer, a species abundant in tropi-

cal Miocene sediments of Jamaica. Benthonic species included *Amphistegina lessonii* d'Orbigny, *Asterigerina carinata* d'Orbigny, and *Trimosina* sp. Large specimens of Peneroplids and Miliolids were also abundant.

An interesting conclusion to be drawn from the presence of such faunas is that the northern Gulf of Mexico must have had a warmer climate than that of the present several times during the Pleistocene. Annual temperatures similar to those of today in the Bahamas and southern Florida were probably approached. These tropical features seem to be related to marine transgressions (Pl. 1) and interglacial stages during which it is postulated that all ice had melted or that there was 50 per cent less ice than now (Trowbridge, 1954, p. 795). Destruction each time of the reef-type organisms may be attributed to drowning in a rising sea rather than to a drastic temperature change.

Conclusions

Transgressive marine wedges in subsurface southeastern Louisiana are seen to correlate with the alluvial formations to the north. If the age assignments as made by Fisk and McFarlan (1955, p. 283) are correct, the Williana transgression is of late Nebraskan and Aftonian age; the Bentley transgression is of late Kansan and Yarmouth age; the Montgomery transgression is of late Illinoian and Sangamon age; and the Prairie transgression would be late Tazewell and Bradyan age. Carbon-14 datings (Horberg, 1955; Fisk and McFarlan, 1955, p. 297) and the sequence of Pleistocene events as summarized by Trowbridge (1954, p. 795) seem to be compatible with such age assignments and the magnitude of the Pleistocene sea-level variations as inferred by foraminiferal evidence.

The predominantly sandy section below the Williana transgression does not offer a means of clearly recognizing the base of the Nebraskan. Both the upper Pliocene and the lowermost Pleistocene at South Pass consist mainly of quartz sands, and both have littoral or supra littoral characteristics as described above. The Pliocene-Pleistocene boundary may be logically assumed, however, on the thickness of overlying intervals to be in the vicinity of 4200 feet in the South Pass well. No evidence of unconformities was seen in the section at South Pass which is near the edge of the continental shelf (Pl. 1), and it is believed that a complete record of Pleistocene history was observed at this position.

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ORIGIN OF PEBBLY MUDSTONES

BY JOHN C. CROWELL

ABSTRACT

Massive marine mudstones containing scattered pebbles crop out at several places in California in strata ranging from Jurassic to Pliocene. Similar rocks, also characterized by unsorted lithology with widely dispersed pebbles, are exposed in Switzerland in Wildflysch strata (Eocene) and in Bavaria in Cenomanian beds.

The California pebbly mudstones are found as lenticular beds in association with slump structures, including slump overfolds. They are interbedded with graywacke, mudstone, conglomerate, and sandstone that were in part laid down by turbidity currents as shown by graded bedding, load casts, current bedding, and other primary structures. Here at places successive stages illustrating the origin of the pebbly mudstones are preserved. Beds of graded conglomerate, laid down on soft water-saturated mud, became unstable, slumped downslope, and mixed pebbles with mud.

Pebbly mudstones originating by this process resemble many rocks designated as tillites. In fact, some glacial episodes may have been introduced into geologic history based on the interpretation of such rocks as tillites without independent supporting evidence. Such "tillites" should therefore be restudied, for they may have originated by slumping and mixing of interbedded gravel and mud penecontemporaneously with deposition.

Small-scale slump structures and mixed rocks, such as those described here, are considered as part of a continuous series which ranges upward from turbidity currents through slumps and large slides to huge tectonic units at the upper extreme. All these phenomena owe their origin to downslope movement under gravity.

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Introduction

Mudstones containing dispersed pebbles crop out at many places in the world and attract attention because of their odd texture, characterized by sparse pebbles set in a mudstone matrix. Although this texture requires that both pebbles and mud be deposited at the same time, it seems unlikely that such poor sorting can come about directly by the simple slackening of current. In this paper pebbly mudstones from four localities in California and two in Europe are described, and their origin is discussed. These outcrops reveal stages in the origin of pebbly mudstones which allow the conclusion that some result from sliding and mixing when unstable gravel layers are laid down by turbidity currents on water-soaked mud. Evidence in support of this conclusion is presented here, and criteria are discussed which may help to discriminate pebbly mudstones formed in this way from those formed by other processes, including glacial processes that result in tillites.

Previous workers have invoked several origins to explain pebbly mudstones. Hohl (1934), Kaiser (1934), and Ackermann (1951) emphasized that these rocks were formed in ways other than by ice rafting or direct wasting from a glacier and so cannot be cited as indicators of glaciation. These workers also stressed that pebbles and matrix cannot have a simultaneous origin since pebbles are deposited from vigorous currents and mud from sluggish currents. Kaiser (1934) deduced that rocks studied by him resulted from the selective deep weathering of conglomerate in place, so that resistant pebbles are left dispersed in an altered clayey groundmass. He could find no evidence that ice-, root-, or organism rafting played a part. Ackermann (1951) favored an origin involving seaward slumping and sliding of coastal clays and near-shore gravels. Such contrasting sediments, mixed during sliding, were thought to be deposited in deep water and came to rest interbedded within fine-grained strata. He cited examples and dealt briefly with factors that played a role in slumping, such as oversteepening of slopes through local heaping of sediments, thixotropy, and other agents possibly responsible for disturbing mass equilibrium.

Submarine slumping as an explanation for shale beds with dispersed clasts has been advanced by many previous workers (Heim, 1908; Lugeon, 1916; Schaffer, 1916; Dorreen, 1951; Kugler, 1953). Such beds, called boulder beds (Dixon and Hudson, 1931), breccia zones (Baldry, 1938), and rubble beds (Dorreen, 1951), contain clasts of both older and contemporaneous rocks embedded in shale with flowage structure. In several areas the evidence is clear that masses of unstable sediment have moved downslope under gravity in a submarine environment. Huge displaced slices, for example, are reported from Venezuela, Peru, and Trinidad (Dorreen, 1951; Kugler, 1953; Renz, *et al.*, 1955; Kamen-Kaye, 1956; Knights, 1956). These and previous workers have emphasized the importance of tectonic activity and earthquake shocks in instigating sliding on a metastable depositional slope. In this paper turbidity currents themselves are advocated as possible additional activators of slumping.

Shales containing exotic blocks and clasts partially resemble pebbly mudstones and have been reported from many parts of the world. In the Alpine Wildflysch they are considered as emplaced by sliding during deposition of encompassing sediment and also as small thrust slices which were milled and worked into incompetent shales (Lugeon, 1916; Ampferer, 1916; Heim, 1921, II, p. 356–361; Cadisch, 1953, p. 182–185). Exotic clasts have also been described from the Ordovician of Newfoundland (Schuchert and Dunbar, 1934, p. 73; Oxley, 1953, p. 30), from the Paleozoic of Quebec (Bailey *et al.*, 1928); from the Carboniferous of southeastern Oklahoma (Miser, 1934, p. 992–1009, west Texas (King, 1938, p. 64–71, 88–92), and England (Dixon and Hudson, 1931; Dixon, 1931), and from the Jurassic (?) and Miocene of California (Woodford, 1925; Page *et al.* 1951, Fig. 16). Other papers on submarine slumping which have been helpful are by Jones (1937; 1939; 1954), Bramlette and Bradley (1940), Fairbridge (1946), and Boswell (1949a; 1949b; 1952). Additional references to slumping are found in the bibliographies of these papers.

Appreciation is here expressed to Edward L. Winterer for many discussions and for reading the manuscript critically. The European localities were visited during the terms of a Guggenheim Fellowship and a Fulbright Award while the writer was on sabbatical leave from the University of California, Los Angeles. Thanks are extended to the Humble Oil and Refining Company for permission to use data on the Grindstone Creek locality. John D. Frick, Alan Jackson, and Michael A. Murphy helped in the field, and Bradford K. Johnson on the mechanical analyses. Field and laboratory expenses in California have been partly defrayed by research grants from the University of California. Appreciation is also offered Betty Bruner Crowell for drafting and Clemens A. Nelson for comments on the manuscript.

Descriptions of Pebbly Mudstones

Grindstone Creek, California

Excellent exposures of marine pebbly mudstone are found along Grindstone Creek, Glenn County, California, about 3½ miles upstream from its confluence with Stony Creek (Fig. 1).

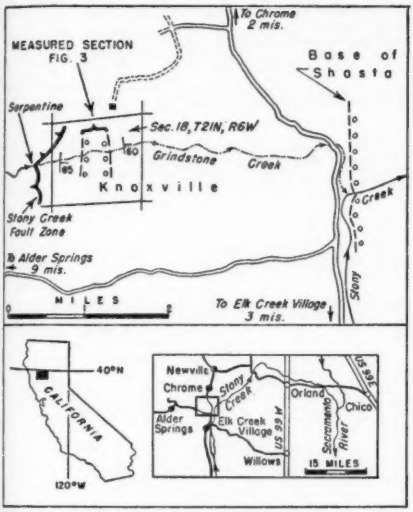

FIGURE 1.—MAPS SHOWING LOCATION OF PEBBLY MUDSTONES AT GRINDSTONE CREEK, CALIFORNIA

The outcrops are in Knoxville beds of the Upper Jurassic, but similar pebbly mudstone is found elsewhere in the region, and within the Lower Cretaceous as well. The stratigraphic section along Grindstone and Stony creeks lies near the center of a belt of remarkably thick Upper Jurassic and Cretaceous strata along the western side of the Sacramento Valley in northern California. Here 32,400 feet of mudstone and graywacke and minor amounts of conglomerate and limestone dips steeply eastward between the Stony Creek fault zone on the west and the valley floor on the east (Fig. 2).

The pebbly mudstones along Grindstone Creek are within the "Grindstone group" of F. M. Anderson's "Knoxville Series" (1945, p. 926), but Anderson has not used the terms group and series according to modern practice. The area is also included on a sketch map by Kirby (1943, Fig. 267), in which he shows the distribution of the Knoxville, Shasta, and Chico groups respectively. Since the stratigraphic nomenclature of these rocks is confused, the column here (Fig. 2) does not indicate what kinds of units are involved. The character, distribution, and origin of the Knoxville strata have been most fully discussed by Taliaferro (1943a; 1943b; 1951) and the biostratigraphy by Anderson (1933; 1938; 1943; 1945). Marine fossils abound locally in the section, and several beds containing *Aucella* [*Buchia*] crop out along Grindstone Creek.

A measured section along Grindstone Creek is shown in Figure 3 and a photograph of part of the section as Figure 1 of Plate 1. Pebbly mudstone is interbedded as lenses as much as 150 feet thick in mudstone, conglomerate, and graywacke. These associated beds are also lenticular and in places display irregular cross-stratification and scour channels. Some graywacke beds are generally only poorly graded; such beds as a whole are poorly sorted, and grading is shown only by slight coarsening at the base and slight decrease in grain size at the top. A few flaggy graywacke beds just above the measured section reveal poor cross-stratification which indicates a local current direction from the northwest, if the beds are considered as turned back to horizontal around the present strike. Similar rocks higher in the section, at the base of the Shasta, display numerous flute casts, groove casts, cross-stratification, graded bedding, load casts, and convolute bedding. Beds in roughly the same stratigraphic position as the pebbly mudstone lenses, but in Salt Creek, 3½ miles south of Grindstone Creek, reveal layer upon layer of current-bedded thin graywacke. The currents from which these rocks were deposited came from the quadrant between north-northeast and west-northwest. Similar small-scale sedimentary structures, reported from other sections and believed to have been laid down by turbidity currents, suggest that the Knoxville strata here were also in part deposited by these agents (Kuenen and Migliorini, 1950; Natland and Kuenen, 1951; Kuenen, 1953a; 1953b; Winterer and Durham, in press; Crowell, 1955; Dzulynski and Radomski, 1955).

The pebbly mudstone along Grindstone Creek consists of about 80 per cent dark-gray indurated mudstone and 20 per cent clasts that are scattered through the matrix and generally do not touch each other (Pl. 2, figs. 1, 2). Although the mudstone in the matrix is mainly massive and featureless, faint rhythmic laminations, slightly graded, can be detected where silt predominates over clay and where very fine sand is present in significant quantities. The weathered mudstone displays subspheroidal fractures and by breaking easily into small shards allows pebbles to roll free. Calcium car-

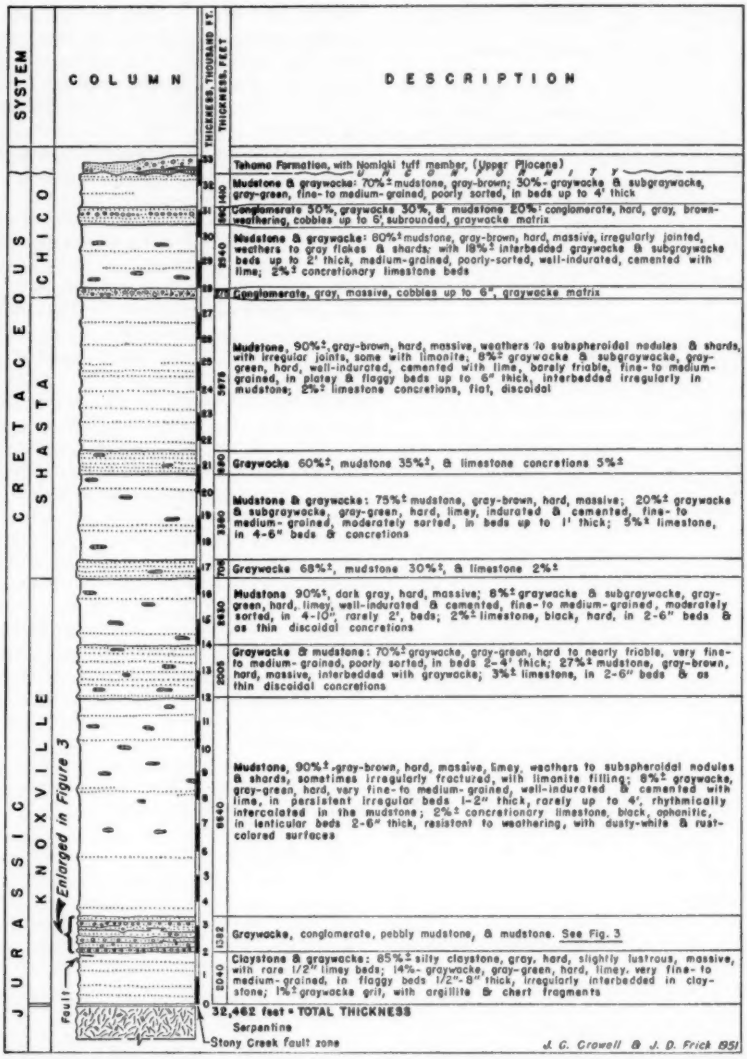
SYSTEM
COLUMN
THICKNESS, THOUSAND FT.
THICKNESS, FEET
DESCRIPTION
JURASSIC
CRETACEOUS
KNOXVILLE
SHASTA
CHICO
Enlarged in Figure 3
Fault
Tehama Formation, with Nomlaki tuff member, (Upper Pliocene)
Mudstone & graywacke: 70%± mudstone, gray-brown; 30%± graywacke & subgraywacke, gray-green, fine- to medium-grained, poorly sorted, in beds up to 4' thick
Conglomerate 30%, graywacke 30%, & mudstone 20%± conglomerate, hard, gray, brown-weathering, cobbles up to 6', subrounded, graywacke matrix
Mudstone & graywacke: 80%± mudstone, gray-brown, hard, massive, irregularly jointed, weathers to gray flakes & shards; with 18%± interbedded graywacke & subgraywacke beds up to 2' thick, medium-grained, poorly-sorted, well-indurated, cemented with lime; 2%± concretionary limestone beds
Conglomerate, gray, massive, cobbles up to 6", graywacke matrix
Mudstone, 90%±, gray-brown, hard, massive, weathers to subspheroidal nodules & shards, with irregular joints, some with limonite; 8%± graywacke & subgraywacke, gray-green, hard, well-indurated, cemented with lime, barely friable, fine- to medium-grained, in platey & flaggy beds up to 6" thick, interbedded irregularly in mudstone; 2%± limestone concretions, flat, discoidal
Graywacke 60%±, mudstone 35%±, & limestone concretions 5%±
Mudstone & graywacke: 75%± mudstone, gray-brown, hard, massive; 20%± graywacke & subgraywacke, gray-green, hard, limey, indurated & cemented, fine- to medium-grained, moderately sorted, in beds up to 1' thick; 5%± limestone, in 4-6" beds & concretions
Graywacke 68%±, mudstone 30%±, & limestone 2%±
Mudstone 90%±, dark gray, hard, massive; 8%± graywacke & subgraywacke, gray-green, hard, limey, well-indurated & cemented, fine- to medium-grained, moderately sorted, in 4-10", rarely 2', beds; 2%± limestone, black, hard, in 2-6" beds & as thin discoidal concretions
Graywacke & mudstone: 70%± graywacke, gray-green, hard to nearly friable, very fine- to medium-grained, poorly sorted, in beds 2-4' thick; 27%± mudstone, gray-brown, hard, massive, interbedded with graywacke; 3%± limestone, in 2-6" beds & as thin discoidal concretions
Mudstone, 90%±, gray-brown, hard, massive, limey, weathers to subspheroidal nodules & shards, sometimes irregularly fractured, with limonite filling; 8%± graywacke, gray-green, hard, very fine- to medium-grained, well-indurated & cemented with lime, in persistent irregular beds 1-2" thick, rarely up to 4', rhythmically intercalated in the mudstone; 2%± concretionary limestone, black, aphanitic, in lenticular beds 2-6" thick, resistant to weathering, with dusty-white & rust-colored surfaces
Graywacke, conglomerate, pebbly mudstone, & mudstone. See Fig. 3
Claystone & graywacke: 85%± silty claystone, gray, hard, slightly lustrous, massive, with rare 1/2" limey beds; 14%± graywacke, gray-green, hard, limey, very fine- to medium-grained, in flaggy beds 1/2"-8" thick, irregularly interbedded in claystone; 1%± graywacke grit, with argillite & chert fragments
32,462 feet • TOTAL THICKNESS
Serpentine
Stony Creek fault zone
J. C. Crowell & J. D. Frick 1951

FIGURE 2.—STRATIGRAPHIC COLUMN OF MESOZOIC STRATA, GRINDSTONE AND STONY CREEKS, CALIFORNIA

bonate and iron oxides are the principal cementing agents, as shown by the white and brown stains in ubiquitous fractures and joints, but the hardness and induration of the musdtone is due predominately to interlocking of angular fragments as the result of compaction.

The size distribution of constituents in the pebbly mudstone is shown in Figure 4, which brings out the distinct bimodal character of the material. Table I shows the combined results of two field pebble counts, 100 clasts in each, of clasts larger than 16 mm: those large enough to be identified easily with a hand lens.

The pebbly mudstone lenses contain numerous contorted blocks of bedded graywacke and other rocks identical in lithology to those stratigraphically below. Some thin-bedded blocks are bent, and mudstone has penetrated between

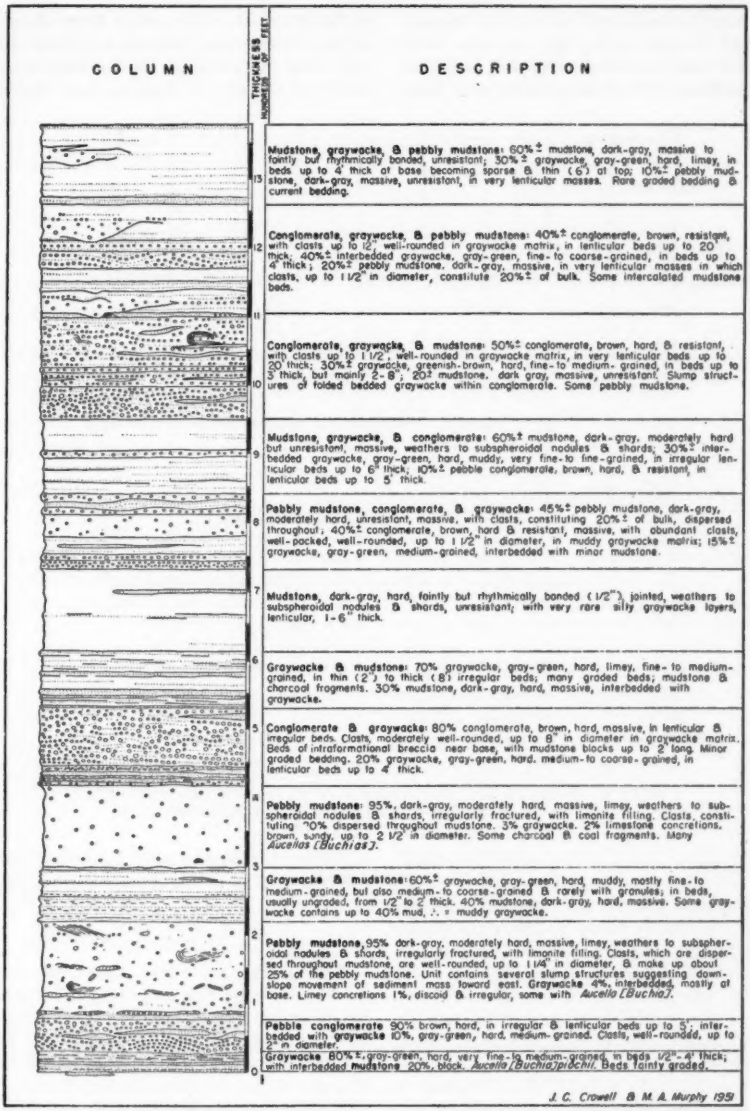
COLUMN
THICKNESS IN HUNDREDS OF FEET
DESCRIPTION
Mudstone, graywacke, & pebbly mudstone: 60%± mudstone, dark-gray, massive to faintly but rhythmically banded, unresistant; 30%± graywacke, gray-green, hard, limey, in beds up to 4' thick at base becoming sparse & thin (6") at top; 10%± pebbly mudstone, dark-gray, massive, unresistant, in very lenticular masses. Rare graded bedding & current bedding.
Conglomerate, graywacke, & pebbly mudstone: 40%± conglomerate, brown, resistant, with clasts up to 12" well-rounded in graywacke matrix, in lenticular beds up to 20' thick; 40%± interbedded graywacke, gray-green, fine- to coarse-grained, in beds up to 4' thick; 20%± pebbly mudstone, dark-gray, massive, in very lenticular masses in which clasts, up to 1 1/2" in diameter, constitute 20%± of bulk. Some intercalated mudstone beds.
Conglomerate, graywacke, & mudstone: 50%± conglomerate, brown, hard, & resistant, with clasts up to 1 1/2" well-rounded in graywacke matrix, in very lenticular beds up to 20' thick; 30%± graywacke, greenish-brown, hard, fine- to medium-grained, in beds up to 3' thick, but mainly 2-8"; 20± mudstone, dark gray, massive, unresistant. Slump structures of folded bedded graywacke within conglomerate. Some pebbly mudstone.
Mudstone, graywacke, & conglomerate: 60%± mudstone, dark-gray, moderately hard but unresistant, massive, weathers to subspheroidal nodules & shards; 30%± interbedded graywacke, gray-green, hard, muddy, very fine- to fine-grained, in irregular lenticular beds up to 6" thick; 10%± pebble conglomerate, brown, hard, & resistant, in lenticular beds up to 5" thick.
Pebbly mudstone, conglomerate, & graywacke: 45%± pebbly mudstone, dark-gray, moderately hard, unresistant, massive, with clasts, constituting 20%± of bulk, dispersed throughout; 40%± conglomerate, brown, hard & resistant, massive, with abundant clasts, well-packed, well-rounded, up to 1 1/2" in diameter, in muddy graywacke matrix; 15%± graywacke, gray-green, medium-grained, interbedded with minor mudstone.
Mudstone, dark-gray, hard, faintly but rhythmically banded (1/2") jointed, weathers to subspheroidal nodules & shards, unresistant; with very rare silty graywacke layers, lenticular, 1-6" thick.
Graywacke & mudstone: 70% graywacke, gray-green, hard, limey, fine- to medium-grained, in thin (2') to thick (8') irregular beds; many graded beds; mudstone & charcoal fragments. 30% mudstone, dark-gray, hard, massive, interbedded with graywacke.
Conglomerate & graywacke: 80% conglomerate, brown, hard, massive, in lenticular & irregular beds. Clasts, moderately well-rounded, up to 8" in diameter in graywacke matrix. Beds of intraformational breccia near base, with mudstone blocks up to 2' long. Minor graded bedding. 20% graywacke, gray-green, hard, medium- to coarse-grained, in lenticular beds up to 4' thick.
Pebbly mudstone: 95%, dark-gray, moderately hard, massive, limey, weathers to subspheroidal nodules & shards, irregularly fractured, with limonite filling. Clasts, constituting 20% dispersed throughout mudstone. 3% graywacke. 2% limestone concretions, brown, sandy, up to 2 1/2" in diameter. Some charcoal & coal fragments. Many Aucellas (Buchias).
Graywacke & mudstone: 60%± graywacke, gray-green, hard, muddy, mostly fine- to medium-grained, but also medium- to coarse-grained & rarely with granules; in beds, usually ungraded, from 1/2" to 2' thick. 40% mudstone, dark-gray, hard, massive. Some graywacke contains up to 40% mud, ∴ = muddy graywacke.
Pebbly mudstone, 95% dark-gray, moderately hard, massive, limey, weathers to subspheroidal nodules & shards, irregularly fractured, with limonite filling. Clasts, which are dispersed throughout mudstone, are well-rounded, up to 1 1/4" in diameter, & make up about 25% of the pebbly mudstone. Unit contains several slump structures suggesting downslope movement of sediment mass toward east. Graywacke 4%, interbedded, mostly at base. Limey concretions 1%, discoid & irregular, some with Aucella (Buchia).
Pebble conglomerate 90% brown, hard, in irregular & lenticular beds up to 5', interbedded with graywacke 10%, gray-green, hard, medium-grained. Clasts, well-rounded, up to 2" in diameter.
Graywacke 80%±, gray-green, hard, very fine- to medium-grained, in beds 1/2"-4' thick; with interbedded mudstone 20%, black. Aucella (Buchia) pallasi. Beds faintly graded.
J. C. Crowell & M. A. Murphy 1951

FIGURE 3.—STRATIGRAPHIC COLUMN OF PART OF KNOXVILLE SECTION CONTAINING PEBBLY MUDSTONE, GRINDSTONE CREEK, CALIFORNIA
The location of the strata shown in this column is indicated on Figures 1 and 2.

layers and forced them apart. At one locality, an angular block of mudstone and fine-grained graywacke, containing an *Aucella* [*Buchia*], is within a pebbly mudstone unit (Pl. 2, fig. 2). These structures and the fossil show clearly that the underlying beds of the Knoxville have contributed clasts to younger units of the same succession. In addition, at several places along the basal contact of a lens, the underlying graywacke layers are twisted and torn and display a stilled view of the brecciation process responsible for the blocks. These pebbly mudstone units

were therefore emplaced as viscous masses, competent enough to pry up beds and twist and break them into fragments, and at several places the loosened beds were folded over when

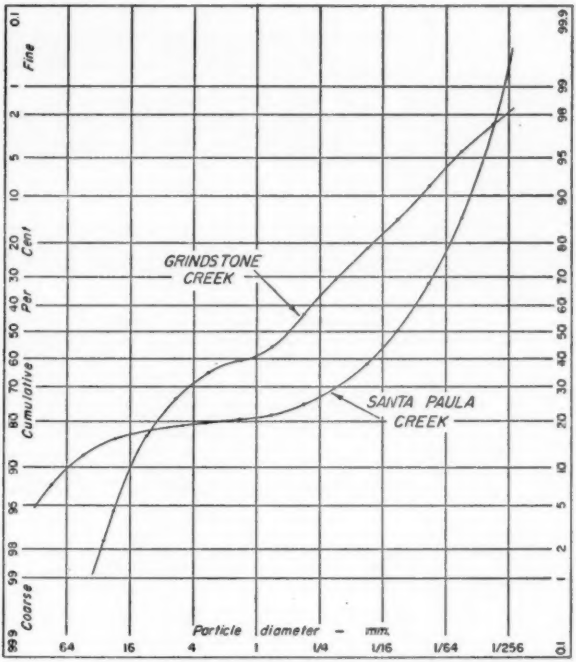

FIGURE 4.—PARTICLE SIZE DISTRIBUTION OF PEBBLY MUDSTONES FROM GRINDSTONE AND SANTA PAULA CREEKS, CALIFORNIA

the strata were still soft and easily deformed. These hooklike structures, here called slump overfolds, show roughly the direction of movement of the muddy mass, which at the places observed is downslope in an easterly and southeasterly sense (Fig. 5; Pl. 1, fig. 2).

Pebbly mudstones along Grindstone Creek therefore were probably formed during downslope movement of viscous masses of mud. Slides plowed into sediments newly laid down on the sea floor and pried loose and broke off fragments which were then kneaded into the moving mass. The mud was viscous enough to support these blocks and pebbles so that they did not sink through the mud and accumulate at the base of the flow. It is not clear from Grindstone Creek exposures, however, whence mud and pebbles were derived and how they came to be mixed together. Since the associated beds display delicate structures and were probably laid down by turbidity currents, the water was probably quiet and therefore fairly deep, and the environment sheltered or relatively distant from shore. In addition, the lenticular shape of conglomerate masses suggests that they were laid down as tongues in channels cut into deposits formed of layers of mud and graywacke.

Santa Paula Creek, California

A pebbly mudstone bed about 2 feet thick is exposed in Santa Paula Creek 3 miles north of Santa Paula, Ventura County, California, and 220 feet north of a steel bridge. The bed is within the lower part of the Pico Formation, Pliocene Series, of the eastward-trending ventura basin, in the section that has been described by Natland and Kuenen (1951). It is interbedded within layers that were laid down by turbidity currents as shown by graded bedding, load casts, small-scale cross-stratification, convolute bedding, pull-aparts, and other structures. Here abundant late Pliocene Foraminifera from associated rocks show that the sea water was from 1000 to 2000 feet deep (M. L. Natland, 1952, personal communication). A cubic foot of

intact pebbly mudstone brought back to the laboratory for mechanical analysis contained a median phalanx of a ruminant as a clast. According to Professor R. A. Stirton, "it compares ... favorably with large individuals of the genus *Odocoileus* (deer)" and is probably late Pliocene in age (Personal communication, May 22, 1952). It seems clear that the deer bone and a pine cone and shallow-water shell fragments also found in the bed were washed into deep water from near shore.

TABLE 1.—PEBBLE COUNT, GRINDSTONE CREEK PEBBLY MUDSTONE

Lithology	Percentage				
	Rounded	Sub-rounded	Sub-angular	Angular	Total
Argillite and chert: gray, white, and reddish; fine-grained, crystalline; hard	20	35	18		73
Mudstone: gray, black, and dark brown; relatively soft	..	..	4	..	4
Graywacke: gray, olive brown; very fine-grained	..	3	3	2	8
Quartz: white, coarse crystalline	..	..	3	1	4
Volcanic rocks: light brown and brown; fine-grained; some with plagioclase phenocrysts	2	5	4	..	11
Total:	22	43	32	3	100

Pebbly mudstone in Santa Paula Creek contains bent strips and angular fragments of sandstone and shale similar in lithology to strata just above and below, as well as clasts of older rocks (Fig. 6). Some pebbles were pressed into sandstone clasts showing that the latter were soft during emplacement. The mudstone matrix, commonly claystone, is massive, homogeneous, and nearly featureless except for faint bands brought out by limonitic stains. The base of the bed is marked by conspicuous load casts where lobes of mud were forced into soft sediment below during deposition. In addition, along the base of the overlying graded bed excellent load casts are displayed, which attest in turn to the plastic character of the pebbly mudstone when the next bed was laid down.

Near-by exposures provide further clues concerning the process responsible for the pebbly mudstone. Within the section are lenticular beds of pebbly gravel and conglomerate, some of which were laid down on layers of mud as much as 15 feet thick. About 100 feet stratigraphically below the pebbly mudstone bed are several thick beds of conglomerate exhibiting large load casts at their contact with underlying shale. Many of these load casts, which contain clasts coarser than in the graded conglomerate above, have deformed rather than truncated the laminations in the mud below and were therefore formed during emplacement of the gravel layer while the mud below was still soft (Fig. 7).

A field sketch showing intraformational breccia and slump structures involving several layers is shown as Figure 8. Irregular synclinal masses of grit and coarse sandstone lie between mushroom-shaped anticlines with disturbed internal laminations. Some silty sandstone layers are pinched and broken and look somewhat like metamorphic boudins. Flaky siltstone clasts are conspicuous at the top of the layer but are also present within the grit and sandstone. At another exposure of the same layer, depicted in Figure 9, disrupted siltstone with broken-off fragments is associated with pebbly mudstone. Pebbles clearly pressed into siltstone suggest that the siltstone was still soft when the pebbly mudstone was emplaced. Near by is another bed which shows a series of complex folds involving several siltstone and shale layers (Fig. 10). Because the tops of these minor folds have been truncated and the next layer passes over them without disturbance it is likely that deformation took place shortly after deposition and before the next bed was laid down.

In summary, data from Santa Paula Creek suggest that the pebbly mudstone layer was formed in water as deep as 2000 feet in association with deposits laid down by turbidity currents. The bottom slope was steep enough so that in places newly deposited layers were unstable and slumped. From time to time vigorous turbidity currents brought gravel and

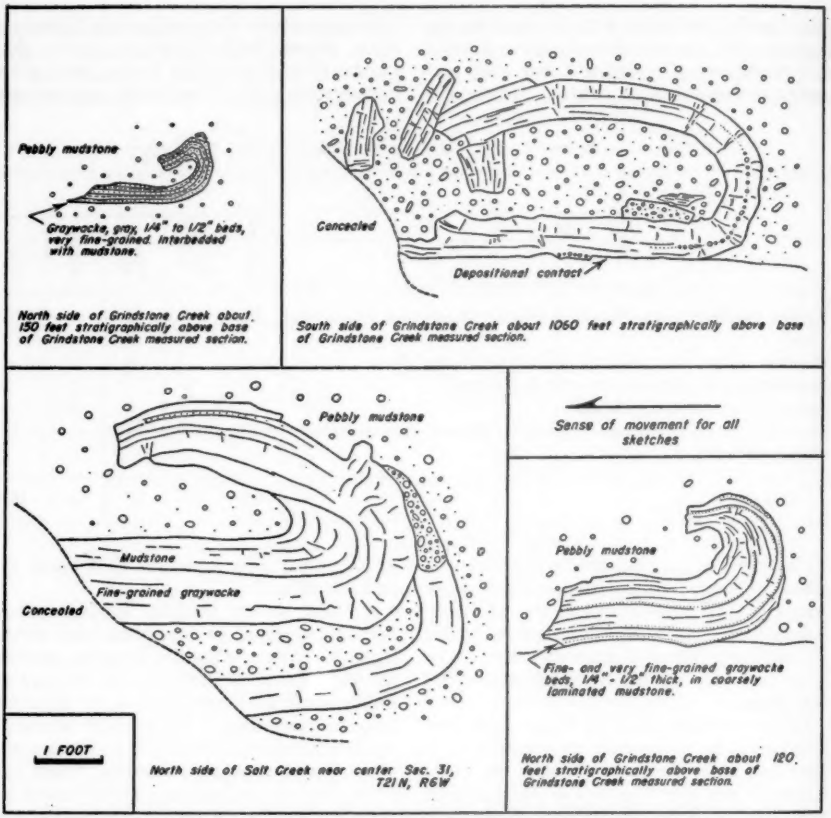
Pebbly mudstone
Graywacke, gray, 1/4" to 1/2" beds, very fine-grained, interbedded with mudstone.
North side of Grindstone Creek about 150 feet stratigraphically above base of Grindstone Creek measured section.
Concealed
Depositional contact
South side of Grindstone Creek about 1060 feet stratigraphically above base of Grindstone Creek measured section.
Pebbly mudstone
Mudstone
Fine-grained graywacke
Concealed
1 FOOT
North side of Salt Creek near center Sec. 31, 7 21 N, R6W
Sense of movement for all sketches
Pebbly mudstone
Fine- and very fine-grained graywacke beds, 1/4" - 1/2" thick, in coarsely laminated mudstone.
North side of Grindstone Creek about 120 feet stratigraphically above base of Grindstone Creek measured section.

FIGURE 5.—SKETCHES OF SLUMP OVERFOLDS AT GRINDSTONE AND SALT CREEKS, CALIFORNIA

sand into the environment and deposited it passively on water-saturated mud. It seems likely, as described in more detail below, that this superposition of gravel on mud was unstable so that downslope movement accompanied by churning ensued, and that during this mixing pebbles were dispersed throughout the mud.

Point Mugu, California

Pebbly mudstone is exposed about 4600 feet northwest of Point Mugu, Ventura County, California, in near-vertical road cuts 15 feet east of culvert A 625 + 72 beneath U. S. Highway 101 Alternate. The strata belong to the lower Topanga Formation, middle Miocene, of the western Santa Monica Mountains (Durrell, 1954). The pebbly mudstone, accompanied by complex slump structures, is interbedded with graywacke and mudstone (Fig. 11). Graded bedding, current bedding, convolute bedding, load casts, intraformational breccia, pull-aparts, and other structures indicate that turbidity currents played a role in sedimentation (Kuenen, 1953a, p. 1063). The orientation of current bedding suggests that much of the sediment came from the south and southwest, if the beds are considered as rotated to horizontal about their present strike.

The Point Mugu pebbly mudstone consists of about 5 per cent clasts mixed with 95 per cent sandy mudstone. Clasts are about half subangular and subrounded quartz, quartzite, mica-schist, volcanic rocks, and half pieces of graywacke, some of which are as much as 6 inches long and are bent and twisted. A discontinuous thin conglomerate layer lies just above the pebbly mudstone (Fig. 11, *cg*). At nearly the same stratigraphic horizon to the west is a series of sheared slump folds with random axes (Fig. 11, sf_1), which may have been formed either before or after a thick over-

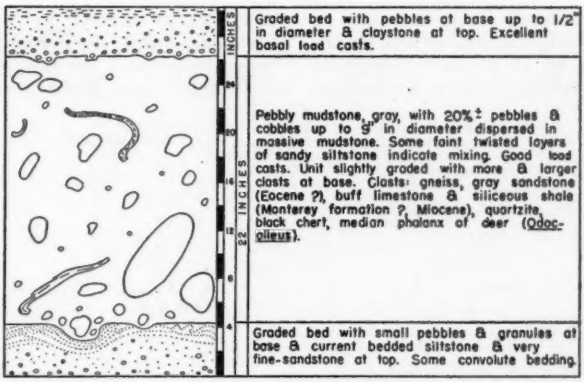
INCHES
Graded bed with pebbles at base up to 1/2" in diameter & claystone at top. Excellent basal load casts.
Pebbly mudstone, gray, with 20%± pebbles & cobbles up to 9" in diameter dispersed in massive mudstone. Some faint twisted layers of sandy siltstone indicate mixing. Good load casts. Unit slightly graded with more & larger clasts at base. Clasts: gneiss, gray sandstone (Eocene ?), buff limestone & siliceous shale (Monterey formation ?, Miocene), quartzite, black chert, median phalanx of deer (Odocoileus).
Graded bed with small pebbles & granules at base & current bedded siltstone & very fine-sandstone at top. Some convolute bedding.

FIGURE 6.—COLUMNAR SECTION OF PEBBLY MUDSTONE BED, SANTA PAULA CREEK, CALIFORNIA

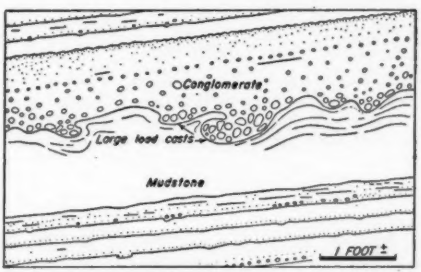

FIGURE 7.—DIAGRAMMATIC SKETCH OF LARGE LOAD CASTS AT BASE OF THICK CONGLOMERATE BED, SANTA PAULA CREEK, CALIFORNIA

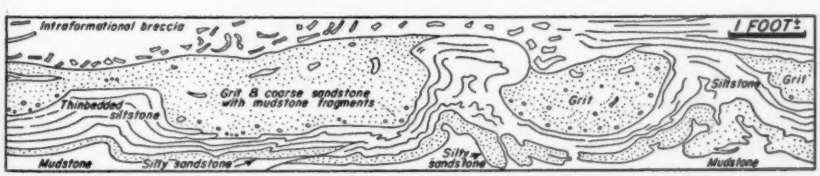

FIGURE 8.—FIELD SKETCH OF SLUMP STRUCTURES AND INTRAFORMATIONAL BRECCIA, SANTA PAULA CREEK, CALIFORNIA

lying sandstone bed was laid down. Some of these folds have been sheared to form lenticular phacoids (Fig. 11, *fd*). At higher horizons and still farther west is another series of slump folds with regular eastward-trending axes (Fig. 11, sf_2, sf_3, sf_4), including some overfolds, which were truncated before thin graywacke layers above were deposited. Beneath these folds are several large load casts, as much as 15 inches in depth, which show that bulbous unstable masses of grit sank into mud and forced water-saturated sediment upward (Fig. 11, *ulc*).

Pigeon Point, California

Upper Cretaceous pebbly mudstone and pebbly sandstone crop out in the sea cliff about 100 yards north of Pigeon Point lighthouse, San Mateo County, California. Here well-rounded polymictic pebbles are mixed with mudstone and fine-grained sandstone in lenticular layers that are interbedded in a thick section of sandstone, conglomerate, and shale assigned to the Chico Formation (Branner *et al.*, 1909). Some exposures show clasts thoroughly mixed with matrix (Pl. 3, fig. 2), others display bent and

irregular silty sandstone blocks and slabs as much as a foot in length intermixed with pebbly sandstone (Pl. 3, fig. 1). At places hard clasts of older rocks have been forcibly pressed into sandstone slabs. These exposures suggest a stilled view of mixing in which churning pebbles and sandy matrix were engulfing soft slabs of newly deposited sandstone.

Les Voëttes, Switzerland

Pebbly mudstone with large angular exotic blocks is exposed in road cuts along the Col du Pillon highway between Le Sépey and Les Diablerets, Vallée des Ormonts, Romande Prealps, Switzerland. The exposures, consisting of Wildflysch facies in Paleocene and Eocene Flysch, are near a large rockfall about 300 m south of the bridge over La Raverette, one of the tributaries of the Grande Eau. The village of Les Voëttes surmounts the bluff above the highway on the east. The tectonic position of these strata is still under debate; some workers place them in Ultrahelvetic nappes and others

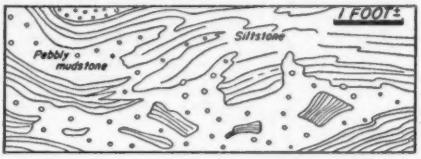

FIGURE 9.—FIELD SKETCH OF PEBBLY MUDSTONE WITH ANGULAR SILTSTONE SHREDS AND BLOCKS, SANTA PAULA CREEK, CALIFORNIA

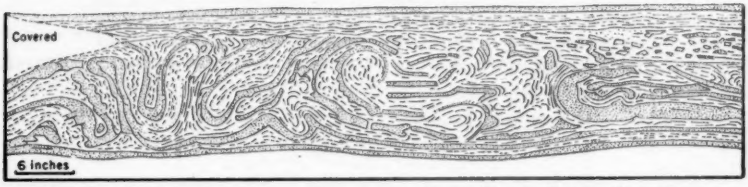

FIGURE 10.—SKETCH OF SLUMP STRUCTURES, SANTA PAULA CREEK, CALIFORNIA

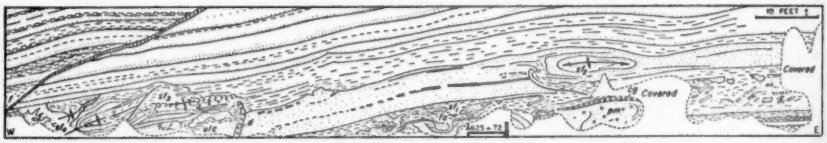

FIGURE 11.—SKETCH OF PEBBLY MUDSTONE AND ASSOCIATED ROCKS AND STRUCTURES, POINT MUGU, CALIFORNIA

Fault (f), slump folds (sf), conglomerate (cg), pebbly mudstone (pm), unstable load casts (ulc), phacoids (fa), and dike (d).

PLATE 1.—PEBBLY MUDSTONE AND SLUMP OVERFOLD, GRINDSTONE AND SALT CREEKS, CALIFORNIA

FIGURE 1.—North bank of Grindstone Creek. (1) Layers of pebbly conglomerate, graywacke, and mudstone. (2) Pebbly mudstone. (3) Pebble conglomerate. (4) Pebble conglomerate and bedded graywacke. (5) Terrace deposits. Located across canyon from units shown between 1050 and 1250 feet in Figure 3. Michael A. Murphy in photograph.

FIGURE 2.—Slump overfold, Salt Creek, California. Folded layers of graywacke within pebbly mudstone. Alan Jackson in photograph.

PLATE 2.—PEBBLY MUDSTONE, GRINDSTONE CREEK, CALIFORNIA

FIGURE 1.—Pebbly mudstone, Grindstone Creek, California

FIGURE 2.—Pebbly mudstone, Grindstone Creek, California. Upper-right quarter of photograph consists for the most part of an angular graywacke clast similar in lithology to graywacke layers below the pebbly mudstone. An *Aucella* [*Buchia*] cast, painted white in the field, is shown above the penny.

FIGURE 1

FIGURE 2

PEBBLY MUDSTONE AND SLUMP OVERFOLD, GRINDSTONE AND SALT CREEKS, CALIFORNIA

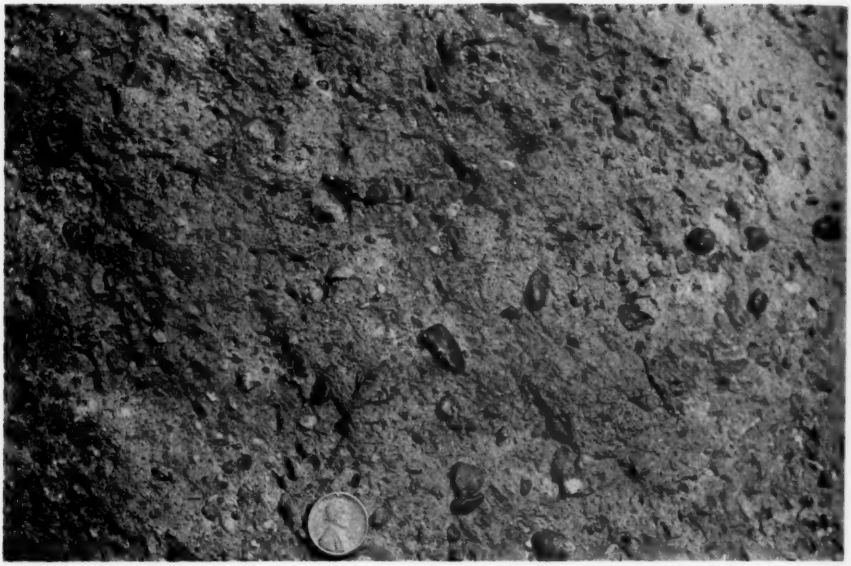

FIGURE 1

FIGURE 2

PEBBLY MUDSTONE, GRINDSTONE CREEK, CALIFORNIA

FIGURE 2

FIGURE 1

VIEWS OF ROCKS, PIGEON POINT, CALIFORNIA

FIGURE 2

FIGURE 1

VIEWS OF ROCKS, LES VOËTTES, SWITZERLAND

in Niesen nappes (Lugeon, 1938, p. 14; Lugeon and Gagnebin, 1941, p. 11; McConnell, 1951; Cadisch, 1953, p. 196–199). A large-scale cross section through the locality has been drawn by Lugeon (1938, Fig. 3) where the Wildflysch under discussion lies in the upper part of his unit 10 near the contact with unit 9.

The Les Voëttes rocks include beds of muddy breccia with angular blocks of granite, as much as 5 m across which are associated with subangular and subrounded clasts of many rock types, including granite, gneiss, and schist. Zones of coarse breccia are indistinctly interbedded with massive pebbly shale (Pl. 4, fig. 2), much of which contains bent slabby clasts of laminated shale (Pl. 4, fig. 1). Large angular oligomictic blocks and contrasting smaller, better-rounded polymictic clasts in places make a roughly graded sequence. The shale matrix, commonly slaty, constitutes up to 50 per cent of the rock. Associated beds of graywacke and shale, although considerably deformed, display graded bedding, convolute bedding, poor load casts, and poor current bedding. Grit and microbreccia beds at many places include numerous chips of shale similar lithologically to that in underlying layers. Complex minor folds and faults within the sequence may have been caused either by slumping or by intense deformation which has obscured evidence of events during deposition. Here again the Les Voëttes exposures suggest that slumping and mixing took place in a region of turbidity current deposition, but where large blocks were contributed from a near-by steep escarpment.

Häuselloch Bach, Bavaria

Pebbly mudstone is exposed in the canyon walls of the Häuselloch Bach, about 3¼ km north-northeast of Hindelang, Bavaria, in the northern Algäuer Alps near the Austro-Bavarian boundary. It is most easily reached by walking southwestward from the automobile road near Unterjoch in the upper Wertach Valley first up the Weissen Bach and then the Häuselloch Bach. These rocks are placed in the Cenomanian of the Upper East Alpine Algäuer nappe (Custodis and Schmidt-Thomé, 1939, p. 407; Custodis, 1951, p. 33–41; Kuhn, 1954, p. 93).

Here again is massive mudstone, generally marly, with dispersed clasts of older rocks as much as 10 cm in diameter and bent slabs similar lithologically to underlying shale. At one place the upper hummocky surface of a pebbly mudstone bed contains broad irregular depressions filled with thin overlapping layers of sandstone and shale. The layers, which suggest deposition on an irregular surface, give way upward to flaggy and "varved" shale, and fine-grained sandstone, grit, and microbreccia. These strata, as well as those near by and underlying the pebbly marl, display graded bedding, current bedding, and some convolute bedding. Slumping in a turbidity current environment was probably responsible for the pebbly mudstone.

Origin of Pebbly Mudstones

General Statement

Pebbly mudstone is used here as a descriptive name for a rock composed of dispersed pebbles in a mudstone matrix without regard to manner of origin. These rocks, characterized by poor sorting and too much matrix to qualify as conglomerate, are formed in several ways, such as: ice and organic rafting, glacier wasting, selective weathering and alteration of a normal conglomerate, deposition from terrestrial mud flows, and slumping. In addition, certain pebbly shales of tectonic origin resemble pebbly mudstones except for marked fissility. In this paper particular attention is paid to slumping in association with turbidity-current deposits, although other origins are discussed briefly. Special emphasis is placed on the recognition of tillites.

Turbidity Currents and Slumping

Stages in the origin of some pebbly mudstones are illustrated above. At Santa Paula Creek,

Plate 3.—VIEWS OF ROCKS, PIGEON POINT, CALIFORNIA

Figure 1.—Pebbly sandstone, Pigeon Point, California. View showing bent clasts of fine sandstone intermixed with pebbles. Scale is 6 inches long.

Figure 2.—Pebbly sandstone, Pigeon Point, California. Scale is 6 inches long.

Plate 4.—VIEWS OF ROCKS, LES VOËTTES, SWITZERLAND

Figure 1.—Wildflysch breccia, Les Voëttes, Switzerland. Detail showing bent slate clasts. Scale in centimeters.

Figure 2.—Wildflysch breccia, Les Voëttes, Switzerland. Large angular blocks of granite and slate on the right grade upward to the left into conglomerate mixed with bent slate clasts. Scale to left of hammer head is 10 cm long.

for example, it seems clear that pebbly mudstone formed only when several factors were associated, as shown by the following description of the process. Within a basin vigorous turbidity currents from time to time deposited gravel on new sediment. If this sediment were sandy and already somewhat indurated, the gravel was laid down on top in a stable layer. On the other hand, if the sediment were soft and water-saturated, gravel sank as great lobes into the substratum. Perhaps the turbidity current had been moving with fair speed across mud, but with decreasing velocity so that coarse material in the flow began to fall out. Up to this moment turbidity within the current was sufficient to maintain a bulk density of the moving mass somewhat less than the bulk density of the underlying mud, and the current moved along the surface of the mud without plowing into it. With a decrease in turbidity as the current slowed down, however, coarse sediment was deposited on the mud surface, and the layer of gravel acquired a bulk density greater than that of the underlying mud. This instability first resulted in the formation of irregular lobes of gravel which at places are seen today as load casts.

If the contrast in bulk densities were great enough in relation to steepness of bottom slope, and the underlying soft layer were thick enough, the lobes sank into the mud and set off downslope movement of gravel and mud together. As this mass gathered speed it churned, mixed, and dispersed pebbles throughout the mud. Perhaps the mass moved basinward as a discrete flow with an incipient internal turbidity, if it were relatively fluid, and so is found today interbedded within strata as a uniform layer of pebbly mudstone. Such a layer displays load casts and crude grading; it has presumably moved far enough from its origin to begin to acquire the characteristics of a new flow. On the other hand, if the mixed mass were viscous, it moved relatively slowly and probably plowed into the substratum. In this way, slabs of underlying sediment were rolled up and incorporated in the mass to form slump overfolds. The flows may have become constricted if they were viscous and pasty and so today are found in rocks as irregular masses and lenses. It seems likely that all gradations exist between fluid, rapidly-moving flows with incipient turbulence, and viscous sluggish slumps.

In such a process no slump or flow will take place unless certain conditions prevail, and the character of the slump or flow will depend largely on the relative interplay of these conditions. The process as pictured here is dependent on (1) the existence of turbidity currents that bring gravel to the site of deposition, (2) the presence of an underlayer of soft mud, which may or may not have been laid down by older turbidity currents, and (3) sufficient bottom inclination so that with instability the mixture of mud and pebbles moves downslope. It seems likely that these three conditions prevailed at several of the localities described above. Structures and textures of sandstone and conglomerate layers, for example, indicate deposition by turbidity currents. These beds are interstratified with siltstone and mudstone layers, some of which are thick enough to provide an unstable foundation. At several places complex slump structures indicate relatively steep bottom slopes.

Sketches and photographs with this paper illustrate stilled views of stages in the slumping process. An early stage, where gravel has been laid down on a thick layer of soft mud so that large unstable load casts have penetrated deeply, is shown in Figure 7. Similar load casts, which have sunk between mushroom-shaped folds, are depicted in Figure 8. Here also silty sandstone layers are in the process of breaking into angular chips and shards to form intraformational breccias. Similar partially mixed rocks are illustrated in Figure 9 and in Figure 1 of Plate 3. The character of the thoroughly mixed pebbly mudstone is shown in Figure 6 and in Figures 1 and 2 of Plate 2 and Figure 2 of Plate 3.

It is assumed that pebbly mudstone will not form unless the slope is steep enough to permit slumping. At the same time, the slope must be gentle enough to allow deposition by turbidity currents and the slow accumulation of fine-grained detritus, so an equilibrium is implied. At one extreme, it is conceivable that the accumulation is metastable and the momentum or "kick" from a vigorous gravel-bearing turbidity current may set off sliding which otherwise would not take place. Perhaps Figure 7, showing unstable lobes of gravel and grit sinking into underlying mud, illustrates a situation where all factors including viscosity were appropriate for downslope mixing except slope. Other factors, such as earthquake shocks, may set off submarine slumps in metastable deposits.

Where the layer of mud beneath gravel has certain characteristics, sliding and mixing are most likely to occur. As stressed above, it must have a bulk density greater than the gravel-bearing turbidity current moving over its sur-

face but lower bulk density than the gravel when deposited. The range of densities possible between these limits is as yet unknown. Here the thickness, viscosity, and thixotropic properties of the layer may be important. Mud with a high content of colloidal clay may have gelled and be essentially solid. With motion, however, the gell structure breaks down, and the material becomes fluid or plastic (Boswell, 1948; Green and Weltmann, 1946; Ackermann, 1949). Under these circumstances initial and sudden overloading from gravel may cause a break through the thixotropic barrier (gel to sol transformation) so that the mass moves and slides easily. Upon coming to rest the mixture again transforms thixotropically (sol to gel), perhaps even quickly, so that slow settling of dispersed heavy clasts is inhibited.

This hypothetical picture of the slumping process, illustrated here by sketches and photographs at different localities, can be proved for any one locality only with difficulty. At Santa Paula Creek, for example, the slumping hypothesis is quite well supported, but at other localities, data are scanty. At several of the localities features associated with pebbly mudstone suggest that they were formed in much the same way as at Santa Paula Creek, and, if so, add a few details to our reconstructed picture. Taken together, information from several localities permits a more detailed reconstruction of the slumping process. In the absence of complete data at each locality, however, it is of course impossible to maintain that the same manner of origin prevailed at each. Such a method of research, though full of possible pitfalls, may be only a little less sound than the practice demanded by poor outcrops. Even at Santa Paula Creek, for example, steps in the slumping process are worked out by assuming that different beds in a closely associated sequence have nearly the same origin. One pebbly mudstone bed cannot be followed far enough to study it both near its upslope origin and its downslope termination. Steps in the process described here, therefore, come from different places and different beds. The reconstruction is accordingly hypothetical and future observations will no doubt add details and modifications when other exposures are studied with slumping in mind.

Tillites

Inasmuch as pebbly mudstones described here look much like some tillites it is desirable to consider criteria to distinguish between them. In fact, some rocks described as tillites in the literature may have been formed by slumping. It is possible that episodes of glaciation have been based on the assumption that mudstone with dispersed clasts could accumulate only as a glacial deposit. The writer felt, for example, on visiting the Squantum "tillite" near Boston on a Geological Society of America excursion in 1952 that these rocks perhaps also formed by slumping and accordingly required re-investigation before their glacial origin could be accepted (Crowell and Winterer, 1953; Sayles, 1914). In addition, E. L. Winterer examined exposures of Gowganda "tillite" of the Subdury district, Ontario, and came to a similar conclusion (Personal communication; Collins, 1925, p. 63–74). Previous workers have also questioned the designation of some pebbly mudstones as tillite (Salomon-Calvi, 1933, p. 105; Ackermann, 1951).

Some tillites, or sedimentary rocks laid down by wasting of glaciers, are characterized by scattered clasts embedded in a mudstone matrix. Many tillite clasts are angular and striated (Pettijohn 1949, p. 212–223; du Toit, 1954, p. 267–276; Campana and Wilson, 1955) and may even have a "flat-iron" shape (von Engeln, 1930, p. 9–16). Other clasts do not display grooves and striae because of granular texture that does not allow their development and preservation, or because they were never scraped against bedrock or neighboring clasts. In addition, even when discovered, grooves and striae on clasts can only be used with caution as indicators of glaciation since straited clasts are found in terrestrial mudflows (Blackwelder, 1930) and in some tectonic environments (Cadisch, 1953, p. 186).

Other criteria may help to distinguish tillites from pebbly mudstones but are also not diagnostic. Thin sections may reveal an abundance of angular rock fragments with an extraordinary range of size grades (Pettijohn, 1949, p. 213), but many mudstones in association with graywackes and microbreccias have nearly the same character. Sandstone and mud sequences laid down by turbidity currents far removed from a glacial environment may be easily confused with varved sequences (Kuenen, 1953a; 1953b). A striated and grooved pavement beneath pebbly mudstone as in the case of the Dwyka tillite may clearly support glacial origin (Coleman, 1926, p. 122; du Toit, 1954, p. 275) but such pavements can even be confused with slickensided fault surfaces (Shrock, 1948, p. 73). Proof of glacial origin of a pebbly mudstone will therefore depend on fairly complete knowledge of both the rock and its geological environment.

Pebbly mudstones described here, such as those at Santa Paula Creek, differ somewhat

from true tillites. Many clasts are better rounded and include shale and sandstone strips which have been torn from underlying layers. Some exposures also display streaks and textural changes formed by mixing. These internal differences, in combination with the characteristics of associated beds, suffice to distinguish the rocks from tillites.

Ice and Root Rafting

Isolated pebbles and cobbles embedded in fine-grained deposits may have been dropped from melting ice floes, from floating roots or kelp holdfasts (Emery and Tschudy, 1941), or rarely by animals (Emery, 1941). Such clasts are sometimes recognized because they have depressed or disturbed laminations in a thin-bedded section. It may be difficult, however, to distinguish them from pebbles scattered along a bedding plane by a vigorous turbidity current which laid down only these clasts at the place in view. Isolated clasts found near the top of graded sandstone layers in turbidity current sequences perhaps were picked up by a late pulse of a flow in the area of flow origin and were not dropped from rafts or ice floes. Such lone clasts may have come along after the beginning of sand deposition, carried by traction instead of suspension, and thus end up in an anomalous position within a graded layer.

Selective Weathering or Alteration

Selective alteration of a conglomerate in which hard and chemically resistant clasts of rocks like quartzite remain unaltered may result in a rock resembling pebbly mudstone (Kaiser, 1934). Clasts of easily-altered types, such as diorite and impure limestone, may decompose to form a clayey groundmass when attacked by ground water during weathering or by hydrothermal solutions. Pebbly mudstones of such origin can probably be recognized by their geologic occurrence and history and perhaps also by vestigial psephitic texture.

Terrestrial Mudflow Deposits

Some lithified terrestrial mudflows consist of pebbly mudstone, but under most circumstances these will be easily recognized since they are interbedded with alluvial fan and related deposits. Rarely mudflows reach playas and are preserved as intercalations in fine-grained lacustrine sequences and may superficially resemble pebbly mudstones described here. Some mudflows acquire rounded polymictic clasts from older conglomerates exposed in the drainage area, but as a rule clasts will be angular and composed of only a few rock types. Striated clasts have been reported from mudflows (Blackwelder, 1930), and from landslides (Kayser, 1921, v.1, p. 466, Fig. 353).

Slide Blocks

Exotic slabs and blocks within a stratigraphic sequence are explained in many areas by submarine sliding. Blocks too large and heavy to be rolled and churned within a pasty slump, such as some of those at Les Voëttes, perhaps slid into position from a rugged upslope source. All gradations probably exist between true turbidity currents at one extreme of a series, through slumps such as those responsible for most of the pebbly mudstones described here, to immense slide blocks. These blocks in turn, composed of either older bedrock or partly consolidated sediment, continue to range in size up to huge masses or structural units resembling tectonic thrust plates at the other extreme.

Immense slices which have moved into a region of deposition have recently been described in Venezuela (Bucher, 1952, p. 55; Renz *et al.*, 1955; Kamen-Kaye, 1956; Knights, 1956). Similar giant slide sheets are also reported from Ecuador (Baldry, 1938; Brown, 1938; Dorreen, 1951) and Trinidad (Kugler, 1953). Breccia zones in which clasts have been rolled with mud during slipping of slide plates cut the beding at a low angle. Such giant slide sheets, which have moved on gentle slopes of mud under the impellment of body forces only, resemble structures described in the northern Appenines where movement has taken place on the *argille scagliose* (Merla, 1951; Maxwell, 1953). In several of these areas the slip layers or breccia zones are composed of slickensided and gougey shale. Clasts dispersed through this matrix, some angular and some rounded and polished, have been milled, rolled, and kneaded into shale during movement. In the Alps these pebbly shales are explained in some localities as the result of tectonic movement in rocks long consolidated, especially where lenses or phacoids of older rocks are found within sheared shale, *e.g.*, Wildflysch, Falknis nappe, Rhätikon, Switzerland (Häfner, 1924; Cadisch, 1953, p. 185). Clasts subject to such milling may display scratches and a high polish and resemble those striated by glaciation (Kayser, 1921, Fig. 224; Cadisch, 1953, p. 186).

A complete series of pebbly mudstones and pebbly shales can therefore be visualized between clasts that have been churned near the

base of a small subaqueous slump and those that have been milled out beneath huge slabs of consolidated rock. Even where the thickness of overburden during sliding was relatively thin, as at Point Mugu, the mudstone matrix can be slightly slickensided and slump folds sheared to form phacoids (Fig. 11 *fa*). Similar structures on a somewhat larger scale are shown by Kugler at Soldado Rock, Trinidad (1938, structure sections), and at many places in the Alps (*e.g.*, Heim, 1921, p. 356–361). At the large extreme, such as in the Appenines and Alps, the size of displaced plates is so great that until recently they were only considered as of tectonic origin. Today, however, owing to workers such as Harrison and Falcon (1936), Brown (1938), Lugeon and Gagnebin (1941), Gignoux (1948), Tercier (1950), van Bemmelen (1950), Goguel (1950), de Sitter (1950, 1954), and Merla (1951) attention is focused on gravitational gliding as a possible explanation for many such structural features.

Summary

This study leads to the conclusion that some pebbly mudstones are formed by slumping in a turbidity-current environment. Gravels laid down on soft mud may become unstable and then slide downslope mixing and churning so that pebbles are thoroughly dispersed throughout the mud. Inasmuch as such deposits resemble tillites some previous workers may have unnecessarily introduced glacial episodes into the geologic record. The study further focuses attention on slumping and sliding as important geologic phenomena. Small-scale slump structures and mixed rocks, such as described here, are pictured as lying near one end of a continuous series which ranges upwards from true turbidity currents through slumps and large slides to immense tectonic units at the upper extreme, and all owe their origin to downslope movement under gravity.

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MARINE GEOLOGY OF THE SOUTHERN HAWAIIAN RIDGE

BY EDWIN L. HAMILTON

ABSTRACT

Since 1950 six expeditions from the U. S. Navy Electronics Laboratory and Scripps Institution of Oceanography have added detailed information to the marine geology of the Hawaiian area.

The Hawaiian Islands are peaks on a great ridge which is flanked, in the southern part, by the Hawaiian Deep beyond which is the Hawaiian Arch. The Deep is about 400 fathoms deeper on the east side. Echo-sounder profiles across the Arch on the east side indicate that it is faulted and more asymmetrical than at first supposed (it is steeper to the west) and has a width of about 200 nautical miles and a relief above the Deep of 300–500 fathoms. Downslope northeast of Oahu the normal structure is interrupted by linear volcanics along the same trends as the Honolulu Series formed by Pleistocene and Recent volcanism on Oahu. The Hawaiian Deep and Arch continue to the northwest past Oahu. Off Molokai and Hawaii the Arch has been faulted.

The sediments off Oahu are mixtures of volcanic sand and silt and planktonic Foraminifera; the latter are dominant down to about 2300 fathoms where the sediment is a clayey silt with relatively large amounts of volcanic sand. The sedimentation off Oahu is apparently controlled by submarine canyons which funnel the material to form delta-like features.

The structure around the southern end of the Hawaiian Islands is due to the response of the earth's crust to the great load of the Hawaiian Ridge; this response is thought to be due mainly to elastic downbowing, but actual foundation failure cannot be excluded from consideration.

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Introduction

The geology of the Hawaiian area has an interest beyond the understanding of the morphology and structure of an island chain, because this archipelago represents a great relatively uneroded load on the crust of the earth. The response of the crust to this load yields valuable information on its structure. In the Pacific Basin, until the past decade, there was little valid information on either topography of the sea floor or on the structure of the oceanic crust. Within the past few years, however, information has been produced at an increasing rate from marine geological explorations and from seismic studies so that a good idea of crustal structure and gross morphology of the sea floor may be deduced.

Surprisingly enough, little is known and little work has been done to date in charting the topography of the sea floor away from the immediate top of the Hawaiian Ridge. The U. S. Coast and Geodetic Survey charts are, as usual, excellent in the coastal and immediately adjacent areas.

The information on which the present study is based was gathered, in large part, since 1950 from six expeditions to and through the Hawaiian area by ships of the U. S. Navy Electronics Laboratory and Scripps Institution of Oceanography. This information was supplemented by data on published unclassified charts of the Coast and Geodetic Survey and the Hydrographic Office. Results of a special survey off southeast Hawaii by Emery (1955) were incorporated into the charts in addition to a special survey off northeast Oahu in 1953 and 1954 (Figs. 1, 2). The survey off Oahu will be discussed in some detail because it was the only area in which sampling and reconnaissance mapping were done.

Most of the soundings of the expeditions were made by the EDO echo-sounder adjusted to a velocity of sound in water of 4800 feet/second. As the result of oceanographic stations in the Hawaiian area the sound-velocity gradient has been determined so that the true depth can be determined from the echo-sounder depth. This information has been interpreted in the form of a curve (Fig. 3) for the convenience of the reader or future investigators in the Hawaiian area; these corrections are valid anywhere in the area of the southern Hawaiian Ridge. The chart of the whole southern part of the Hawaiian Ridge (Fig. 1) is based on echo-sounder depths; that of the area off Oahu (Fig. 2) shows true depths.

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General

The Hawaiian Ridge is a tremendous volcanic feature about 1600 miles long; at the southern end it shows the greatest continuous relief, from submarine deep to subaerial mountain top, in the world (about 32,000 feet: Dietz and Menard, 1953, p. 103). This ridge appears to rise from a low swell of the ocean floor, and its topmost peaks project above the water to form the basaltic islands. The volcanic activity has long been known to have progressed southeastward; the present volcanic activity is confined to the island of Hawaii. The southern Hawaiian Ridge and its islands are believed to have been formed during the Tertiary. The island of Hawaii may have been formed as late as the Pliocene (Stearns and Macdonald, 1946, p. 183). The general physiography of the Hawaiian Islands is concisely summarized by Stearns (1946) and it is unnecessary to repeat it here.

The first detailed discussion of the Hawaiian Deep, Arch, and Swell was by Dietz and Menard (1953). They noted that the Hawaiian Ridge is superposed on a low swell of the ocean floor and that at the southern end of the Ridge is a deep along the eastern side which hooks around the south side of the island of Hawaii. On the eastern side of the Deep is a broad arch of the sea floor. Dietz and Menard (p. 102) note that this arch has a width of about 220 nautical miles off Oahu and 210 miles off southeast Hawaii and is asymmetrical with the gentler gradient on the side away from the islands. Off Oahu they record a relief of 3150 feet (525 fathoms) from the bottom of the deep to the top of the arch, whereas off southeast Hawaii this relief is 2100 feet (350 fathoms). Since the 1950 Mid-Pac Expedition there have been several crossings of the Hawaiian Deep and Arch, so that the picture can now be corrected in detail and extended.

Geomorphology

General Structure

The latest interpretation of the generalized structure (Fig. 4) shows the Hawaiian Ridge,

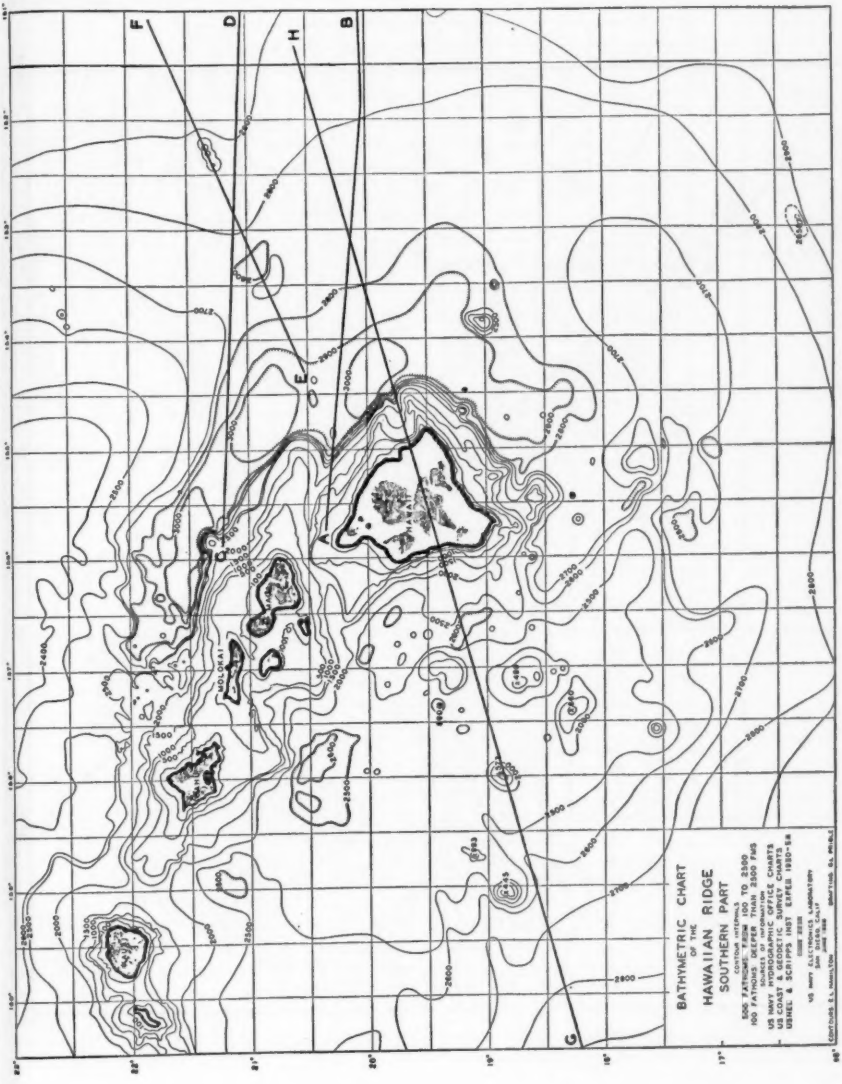

FIGURE 1.—BATHYMETRIC CHART OF THE VICINITY OF THE SOUTHERN END OF THE HAWAIIAN RIDGE

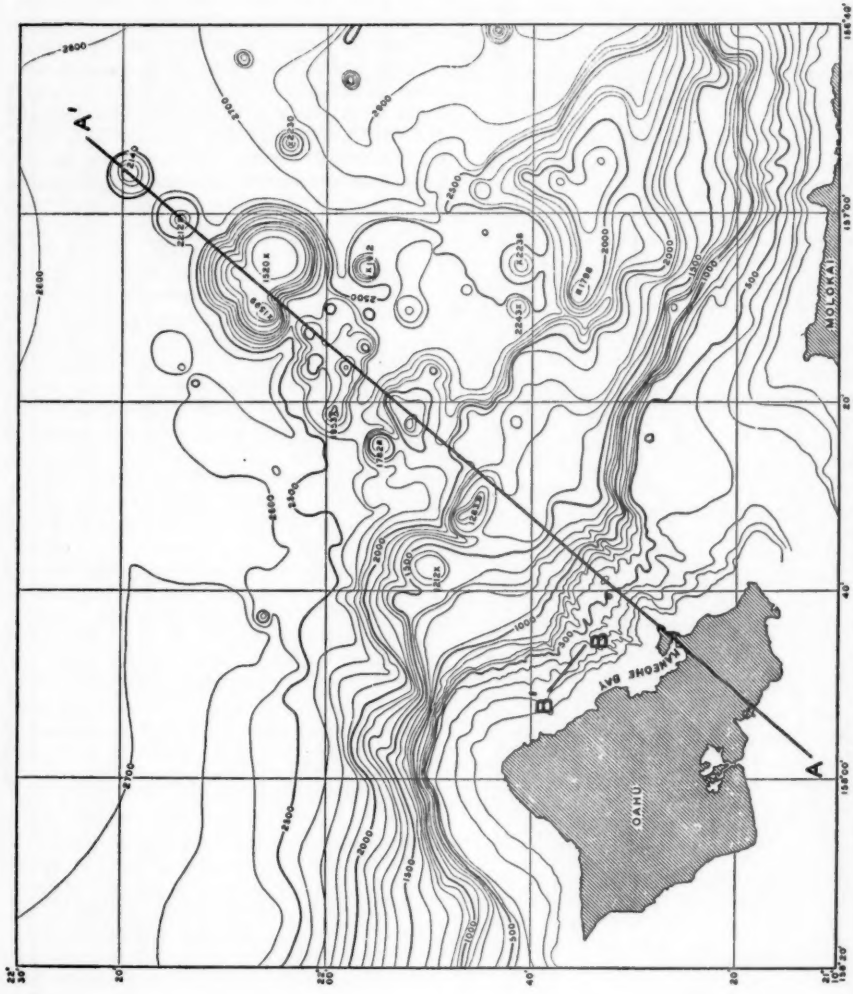

FIGURE 2.—BATHYMETRIC CHART OF AN AREA NORTHEAST OF OAHU; TRUE DEPTH CONTOUR INTERVAL: 100 FATHOMS

its topmost peaks above water, occupying a central and relatively depressed area within the Hawaiian Deep and Arch. On the northeast between the Ridge and the Arch. To the northeast and north of Oahu the Deep is outlined by the 2700-fathom (true depth) contour line.

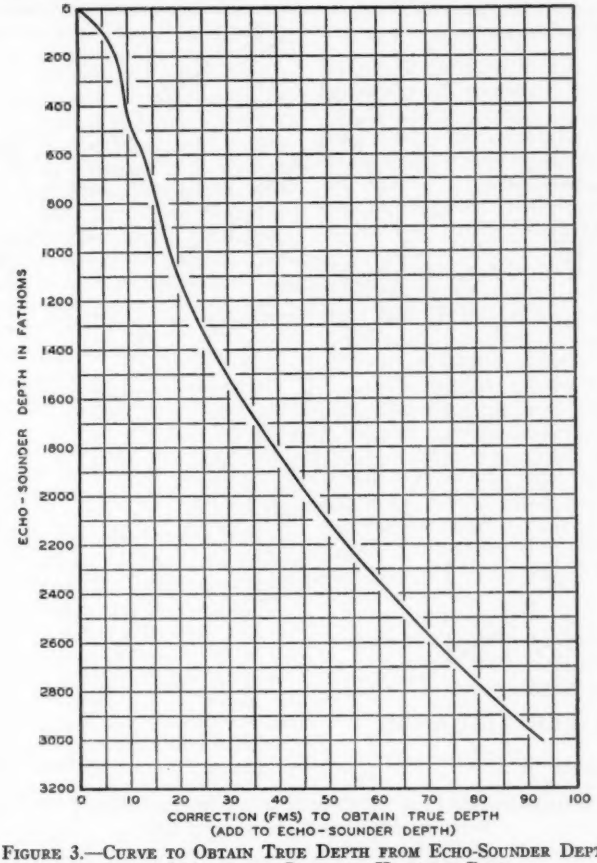

FIGURE 3.—CURVE TO OBTAIN TRUE DEPTH FROM ECHO-SOUNDER DEPTH IN THE AREA OF THE SOUTHERN HAWAIIAN RIDGE

FIGURE 2.—BATHYMETRIC CHART OF AN AREA NORTHEAST OF OAHU; TRUE DEPTH CONTOUR INTERVAL: 100 FATHOMS

the Hawaiian Deep and Arch extend well past Oahu to the northwest where their exact extent and configuration are uncertain. Both the Deep and the Arch swing around the southern end of the Island of Hawaii and can be traced to an area southwest of Oahu. The several crossings of the Hawaiian Arch in 1953 and 1954 indicate that it is more asymmetrical than first supposed, and that the axis and outer limit of the Arch are much closer to the island chain.

Hawaiian Deep

The Hawaiian Deep lies adjacent to the central ridge and is formed by the depression East of Oahu, Maui, and Hawaii, the Deep is below 3000 fathoms. Around the south end of the island of Hawaii the Deep grows progressively shallower until, on the southwest side of Hawaii to an area south of Oahu, it is represented by the 2500-fathom contour line; a few basins are below 2600 fathoms. Beyond Necker Island the Deep on the southwest side is indicated by the 2500- and 2600-fathom contour lines (Dietz and Menard, 1953, Fig. 1). Most of the Deep is shallower than the general level of the ocean floor outside the Arch (2900 fathoms ±) except for the area between Oahu and Hawaii on the northeast side of the ridge, where the depth is below 3000 fathoms.

The special survey off northeast Oahu in 1953 revealed a number of volcanic submarine peaks in the Hawaiian Deep (Figs. 2, 4). The largest of these seamounts, with a basal dimension of 10 by 15 nautical miles and height of 6600 feet, is situated almost in the axis of the Deep.

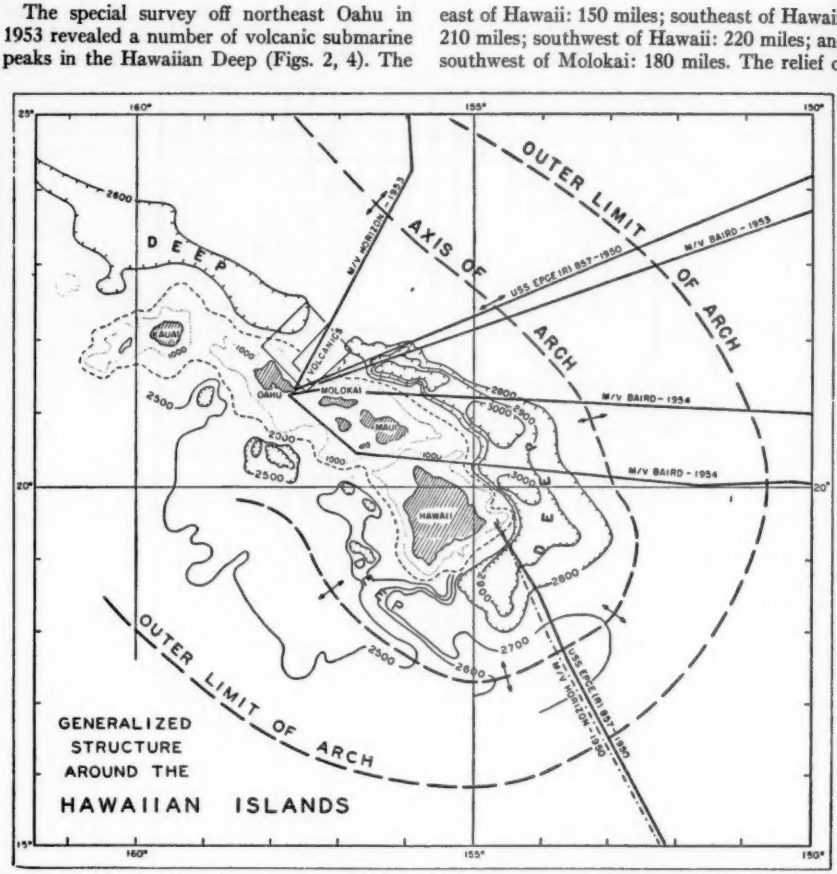

FIGURE 4.—GENERALIZED STRUCTURE IN THE VICINITY OF THE SOUTHERN END OF THE HAWAIIAN RIDGE
Rectangle northeast of Oahu is area of survey.

Hawaiian Arch

The Arch (in profile) starts at the axis of the Deep and is defined on the outside (away from the islands) by the 2900-fathom contour; this contour sweeps in a great horseshoe bend from northeast of Oahu around the south end of the Ridge to an area southwest of Oahu. The wave-like configuration of the Arch can be readily seen in the actual echograms in Plate 1.

The width of the Arch is about as follows: northeast of Oahu: 240 nautical miles; northeast of Hawaii: 150 miles; southeast of Hawaii: 210 miles; southwest of Hawaii: 220 miles; and southwest of Molokai: 180 miles. The relief of the Arch above the Deep varies from 1800 to 2400 feet, except in an area northeast of Molokai where it is as much as 3600 feet.

Continuous echograms of the Arch (Pl. 1) illustrate its asymmetry. On the flank toward the islands the slope is about 0°15′; on the outside of the Arch the slope is about 0°06′. Study of the chart of the whole area (Fig. 1) shows that the crest of the Arch northeast of Oahu is about 2400 feet higher than the crest off northeast Hawaii. The Arch from southeast of Hawaii around to the southwest of Maui grows progressively shallower (from about 2700 to about 2300 fathoms).

The smooth profile of most of the Arch, espe-

cially on the side toward the islands, indicates that aprons of sediment or lava flows had begun to spread outward from the islands before the Arch and Deep were formed (Menard, 1956, p. 2199). A study of the echograms across the Arch (Pl. 1) reveals relief due to small hills on the order of 900-1500 feet. All the profiles show faulting of the Arch. The fault in the top of the profile northeast of Hawaii (Pl. 1, E-F) is almost a classic textbook example of a "keystone" or gravity fault; the fault on the west side of the crest has a displacement of about 600 feet, and the whole top of the Arch has apparently been dropped down. Much of the relief seems to be connected with these faults and is probably the result of magmatic effusion along the fault planes.

Marine Geology of an Area Northeast of Oahu

General

The island of Oahu resulted from the coalescence of two great volcanoes which, according to Stearns (1946, p. 73), were formerly separated by water. The older of these two volcanoes is the Waianae Range on the western side of the island and the younger is the Koolau Range on the eastern side. Both volcanoes are thought to have been formed in Tertiary time. At present they are deeply eroded, and the Koolau volcanic rocks overlie, in places, the soil-covered Waianae lavas. Schofield Plateau connects the two ranges.

During late Tertiary and Pleistocene times the two ranges were eroded to about their present form. After the island was sculptured by erosion to about its present form it underwent several emergences and submergences; at the present it is thought to be submerged about 1200 feet below a previous well-defined shore line (Stearns, 1946, p. 78).

Dietz (Personal communication) believes that submergence is the dominant movement and that emergences are due for the most part to eustatic changes.

The last eruptions of volcanic materials on Oahu occurred during late Pleistocene and early Recent times across the southeast end of the Koolau Range, the present southeast end of the island. These eruptions produced many lava flows and cinder cones and most of the distinctive topographic features of that end of the island, such as Diamond Head, the Punchbowl, Koko Head, and the islands just off the coast on the southeast side. These basalts and pyroclastic rocks have been named the Honolulu Series. Most of these late eruptions were along definite southwesterly trends, or rifts, which, in general, cut across the Koolau Range.

Geomorphology

Figure 2 illustrates for the first time the topography off northeast Oahu which is, in general structure, similar to that of the island chain to the south: Oahu is a great peak on the Hawaiian Ridge; the Deep is to the northeast and parallel to the ridge, and the Hawaiian Arch is beyond the Deep. Off northeast Oahu the Hawaiian Deep is well defined by the 2700-fathom contour line (Figure 2), except for the volcanic peaks which lie downslope from Oahu. The top of the Hawaiian Arch in this area is about 1800 feet above the bottom of the Deep.

The numerous seamounts downslope northeast of Oahu are illustrated in plan in Figure 2 and in profile along traverse line *A-A'* in Figure 5. The volcanic character of these seamounts is indicated by their slopes, symmetrical profiles, oval shapes in plan, their association with volcanic features of the Hawaiian Ridge, and by the volcanic material cored from their slopes or the adjacent sea floor. On one of the seamounts olivine basalt pebbles were brought up in a core from halfway down the seaward slope.

Some of the side slopes of the cones and of the submarine slopes of the Ridge off Oahu compared with similar features off Hawaii (Emery, 1955) are as follows:

Feature	*Side Slopes*		
	max.	*min.*	*average*
Seamounts off northeast Oahu	18°	10°	14°
Seamounts off east Hawaii	...	...	17°
Submarine slopes off northeast Oahu (three slopes)	...	...	7°
	6°	4°	5°
	17°	4°	9°
Submarine slopes off east Hawaii	...	...	11°
Subaerial slope of Mauna Loa	11°	3°	7°

These submarine slopes are uncorrected, but the correction would ordinarily be less than 1°.

Most of the volcanic cones downslope from the island of Oahu are probably adventive cones on the side of the main volcano (Koolau) which formed the east side of the island. The large cone in the center of the Hawaiian Deep, however, seems to be an independent cone. This large feature is about 6600 feet higher than the surrounding ocean bottom and has a basal diam-

eter of 10–15 nautical miles. The side slopes on this particular feature are 12° on one side and 15° on the other (Fig. 5).

On the northeast side of Oahu between Kahana Bay and the southeast tip of the island are several deep gullies which eventually, on the lower slopes, deepen and can be called canyons (Fig. 6; Fig. 2, B-B′). Figure 6, which is a tracing from the EDO echo-sounder record, indicates how marked this gullying is off Kahana Bay. These gullies are 250–500 feet deep. Off Kaneohe Bay and Kaneohe Peninsula the gullies deepen in a short distance to more than 600 feet and widen at the top to about a nautical mile. These gullies and canyons lead, at depths between 1000 and 1500 fathoms, to topography which seems to be a sediment fan which is in part ponded behind volcanic peaks farther offshore (Fig. 2). Part of this fanlike configuration could be formed by lava extruded from the flank of the Koolau volcano.

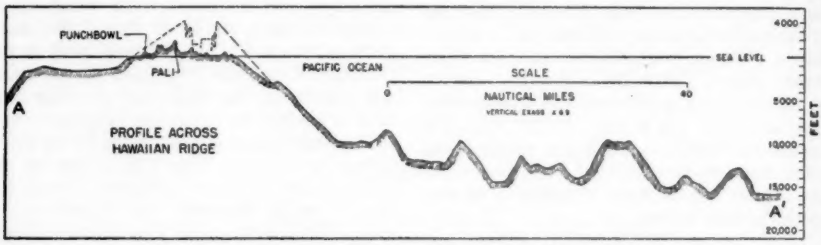

FIGURE 5.—PROFILE ACROSS THE SOUTHEAST END OF OAHU
See Figure 2 for location of the profile line.

Sediments

Fourteen gravity cores 2 inches in diameter were taken in a fanlike sampling pattern downslope from Oahu during the 1953 expedition. These cores ranged in length from a few inches to 5 feet. For the sake of brevity the sediment distribution and properties of the sediments will be combined. Station locations are classified by the Navy.

Down the northeast submarine slope of the island of Oahu, in depths more than 300 fathoms, are three general types of sediments: (1) sandy globigerina ooze, (2) clayey silt, and (3) deep-sea red clay.

Near the island at depths between 300 and about 2100 fathoms the dominant sediment is a sandy globigerina ooze composed for the most part of planktonic Foraminifera, minor amounts of mollusk fragments, echinoid spines, and a few grains which are minerals and rock fragments from a basaltic source. The calcium carbonate content of this sediment is for the most part in the form of the shells, or tests, of planktonic Foraminifera and varies from 34 to 78 per cent by weight. About 20 per cent of this sediment is in the sand size (median grain diameters between 2 and 0.062 mm), but again most of these "grains" are planktonic Foraminifera; wet densities are between 1.46 and 1.56 gr/cc and porosites from 60 to 70 per cent.

The clay fraction (grains less than 4 microns in diameter) of these sandy globigerina oozes averages about 30 per cent.

Farther offshore and downslope in depths from 2100 to 2600 fathoms the dominant sediment is a clayey silt. This material has a median diameter of about 11 microns and about 12 per cent is in the sand-size range. The calcium carbonate content is about 16 per cent; wet densities are between 1.28 and 1.58 gr/cc and porosities 60–80 per cent. This sediment, in the sand sizes, is a mixture of volcanic rock fragments and minerals from a volcanic source and planktonic Foraminifera. In most places the volcanic detrital material is dominant. In marine geologic literature this type of sediment has been called a "volcanic mud", but the writer prefers not to use the generalized term of "mud".

The third general type of sediment off the northeast side of Oahu is the deep-sea red clay. This material, off Oahu, has a median grain diameter of 2 microns. Less than 1 per cent is calcium carbonate, and less than 1 per cent is sand-sized grains. Four cores had an average of 69 per cent of grains in the clay size. Wet densities were between 1.38 and 1.49 gr/cc and porosities were 70–80 per cent. Two cores were taken on the southwest flank of the Hawaiian Arch off northeast Oahu and one just beyond the crest of the Arch to the northeast. All these cores were composed of red clay and minor

amounts of sand-sized volcanic mineral grains and rock fragments. The shallowest red-clay core (2326 fathoms) was taken near the crest of the Arch. Inasmuch as this is about the shallowest portion of the Arch in the southern Hawaiian

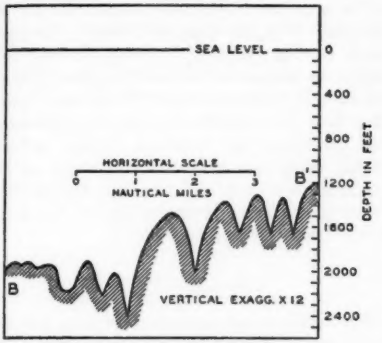

FIGURE 6.—GULLIED SLOPE OFF KAHANA BAY, OAHU
See Figure 2 for location of the profile line.

Ridge area (with the exception of seamounts) most of the Arch is probably covered by red clay. The tops and side slopes of seamounts on the Arch at depths less than about 2300 fathoms are probably covered with globigerina ooze.

Conclusions

An examination of the volcanic features off Kaneohe in an area surveyed on a 5-mile grid reveals a definite lineation in a northeasterly direction. As detailed by Stearns (1946) the late volcanic eruptions in the southeast part of Oahu, which formed the Honolulu Series of volcanic rocks and pyroclastics, trended definitely in the same northeasterly direction. A comparison of the two trends on the same chart reveals the distinct possibility that these submarine features were formed at the same time (Pleistocene and Recent) and are part of the same general eruptive trend. Macdonald (Personal communication) believes that these peaks are along the main secondary trend direction in the islands.

There are four general sources of the bottom sediments off northeast Oahu; (1) volcanic materials from the island, (2) volcanic material from the submarine eruptions which formed the seamounts, (3) planktonic deposition from the water mass which includes both the organisms and the finely divided sediment from other locations, and (4) the material such as shell fragments and echinoid spines provided from bottom-living animals.

The agencies which distribute the sediments from the above sources are bottom currents, slumps, and turbidity currents, which transport sediment along the bottom, the "Kona" winds from the south which carry fine material from nearby islands, and the normal movements of the upper water masses from which the planktonic tests and suspended sediments fall.

Extending around most of the eroded islands and former islands of the Pacific is an apron of sediments which smooths the normally roughened topography of the sea floor (Menard, 1956, p. 2195). This apron of sediment is everywhere around the Hawaiian Ridge except in the area downslope northeast of Oahu. Here, on the normal sediment apron, extensive volcanism formed the many seamounts off northeast Oahu. These seamounts have helped control the sediments moving downslope from the island of Oahu by causing a certain amount of sediment ponding on the island side of the seamounts (Fig. 2).

The gullies and submarine canyons mentioned previously have also aided in the control of sedimentation by apparently funneling the detrital material from the island through the canyons so that they arrive in specific localities to form fan and deltalike deposits which alter the shape of the bottom.

An examination of Figure 2 reveals that the submarine canyons off Kaneohe lead directly down to fan and deltalike features formed just below the 1100-fathom contour line to the north of Kahana Bay and below the 1500-fathom contour line to the south off Kaneohe. Although the island of Oahu is thought to have been elevated and submerged several times during its geologic history the present depth of the mouths of the canyons are below any postulated previous estimates of submergence.

The end result of the movement of sediments from their sources and the topographic control of the resulting movements is that near shore off northeast Oahu, in depths between 300 and 2100 fathoms, the bottom deposits are dominantly formed by the rain of planktonic Foraminifera from the waters above and minor amounts of sand- and silt-sized particles of volcanic rocks and minerals, mollusk-shell fragments, and echinoid spines. The rain of planktonic materials is so great as to mask the other constituents. Farther offshore the same amount of material is descending, but solution of the calcium carbonate tests of the planktonic material takes place in the water column and on

the bottom; $CaCO_3$ is, therefore, a minor constituent of the sediment. The result of this diminution of planktonic Foraminifera on the bottom is that the other constituents form the mass of the sediment. This results in the deposition of a clayey silt in depths from 2100 to 2600 fathoms, except on the Arch, in this particular area. Emery *et al.* (1954, p. 73) find that the general lower limit of deposition of globigerina ooze in the northern Marshall Islands area is about 2400 fathoms. Even farther offshore, and deeper, where the calcareous planktonic material is almost entirely absent the normal sediment is red clay. In the area off Oahu this red clay has anomalous amounts of sand-sized particles of volcanic material. A core taken just below the slopes of the great seamount in the Deep off Oahu is a good example of the interaction of these factors. In this core there is a high content of sand-sized particles from volcanic sources, so that the material forms a clayey silt even though in the Hawaiian Deep at a depth of 2567 fathoms a deep-sea red clay with rare sand-sized particles would be expected. The source of this sand is probably the great seamount which lies just to the east.

Discussion and Conclusions

General

A number of papers have been written on the Hawaiian structure (*e.g.*, Chubb, 1934; Betz and Hess, 1942; Gunn, 1943; Stearns, 1946; Dietz and Menard, 1953). The most recent paper which details a realistic picture of the structure is that by Dietz and Menard who describe a swell formed by a broad uplift of the sea floor, possibly due to subcrustal currents, followed by volcanism, faulting, and then by regional and local isostatic subsidence. In a recent paper Hess (1954) discusses oceanic ridges in general and speculates on their origin by serpentization of peridotite below the Mohorovičić discontinuity. It should be noted that the depressions noted by Betz and Hess along the outside of the Hawaiian Swell are not the Hawaiian Deep, which is very near the sides of the ridge.

The ridge of the Mid-Pacific Mountains, a chain of islands in Cretaceous time, intersects the Hawaiian Ridge at Necker Island (Hamilton, 1956). That these two structures are possible parts of the great fracture zones of the northeastern Pacific is indicated by Menard (1955, p. 1163, 1166, 1196). The writer is in agreement with this opinion and believes that all these great linear features in the North Pacific are best explained as part of one great integrated system.

New light is thrown on the structure and genesis of the Hawaiian Ridge, Deep and Arch, because of the evidence which has been gathered in the area since 1949. Any hypothesis for the Hawaiian structure must now explain the great linear central ridge, with marginal deeps and bulges on each side, the fact that the Deep on the northeast side is about 300 fathoms deeper than that on the southwest side, the asymmetry and faulting of the bulges or Arches, and the greater depth of the crest of the Arch on the southeast side, plus the fact that the greatest depth of the Deep, is almost that of the ocean floor outside the Arch.

Hawaiian Swell

The southern part of the Hawaiian Swell is made up of the central Hawaiian Ridge and two bulges of the sea floor on each side (Figs. 1, 7). The 600-mile wide Hawaiian Swell of Betz and Hess (1942) is a simplification of the overall structure, but was, of course, based on the meager data available at that time. It has always been assumed that the oldest peaks of the Hawaiian Ridge appeared above water at the northern end and that the peaks (islands) to the south are progressively younger. This assumption is based on the progressive waning of volcanism to the south and to the erosion of the northern peaks to flat banks. The marginal deeps on each side of the ridge are additional evidence of progressive youthfulness to the south; most of the northern marginal depressions have been filled so that the northern part of the Ridge appears in cross section as a swell whereas the southern depressions are at present unfilled. Menard (1956) has discussed the moating around Pacific islands and the filling of these moats to form "archipelagic aprons". Menard compares the filled depression around the Marquesas with the Hawaiian unfilled depressions and suggests that the Hawaiian Deep will eventually be filled by volcanic material and sediments to an equilibrium profile to the top of the Arch; apparently this has happened in the northern part of the Hawaiian Ridge.

Menard (1955, p. 1183) in his discussion of the great fracture zones of the northeastern Pacific has shown that most of these zones are huge raised welts on the ocean floor. If volcanic materials had been extruded from the central parts of these welts and had filled the central

depressions the morphology of the zone would probably be that of a line of seamounts atop a great swell of the ocean floor. A line of seamounts with side bulges would also appear as a swell if the marginal deeps are disregarded. If these swell and seamount relations are true deliver valid information on which to build concepts. The young science of marine geology also needed to deliver valid information on the morphology of the ocean floor before concepts involving the whole earth could be studied. One of the first major works on the strength and

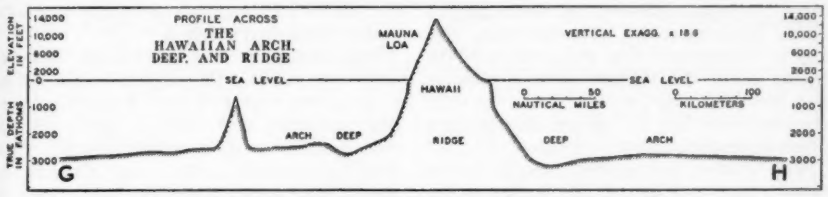

FIGURE 7.—PROFILE ACROSS THE HAWAIIAN ARCH, DEEP, AND RIDGE
For location of this profile see Figure 1.

then *ad hoc* sub-crustal convection currents are not necessary to explain formation of each of the great swells of the Pacific Basin on which seamounts are often linearly arranged; they were formed rather by plastic yielding of the crust along great fracture zones of the type envisaged by Menard (1955) for the northeast Pacific.

Hawaiian Arch and Deep

General.—The Hawaiian Arch has been described above as extending from northeast of Oahu, all the way around the south end of Hawaii to a position southwest of Oahu. The Deep is the result of the arching, or build-up of the sea floor and downbowing of the crust under the Ridge. The position of the Ridge within the Deep and Arch is positive proof that the crust of the earth has yielded to the great load of the Ridge. The formation of the Arch, then, is due to this yielding and is a surface expression of the response of the crust to the subsidence of the ridge and/or crustal yielding. There are at least two ways in which the crustal bulges along the Ridge could have been formed without special subcrustal convection currents beneath the Hawaiian Ridge: (1) the bulges are the result of the elastic response of the crust to a great load, or (2) the crustal foundation below the Hawaiian Ridge has plastically failed because of stresses beyond its ultimate bearing capacity.

Strength and structure of the earth's crust in the Pacific.—The structure and strength of the earth's crust has long been a matter of major interest in geologic research. Most of the ideas had to be considered speculative until the young sciences of geophysics and seismology could structure of the earth was the study by Barrel (1915) who postulated the structure and strength from geodetic and theoretical considerations. The modern concepts are well summarized in the series of papers in Gutenberg, *et al.* (1952), in the symposium papers edited by Poldervaart (1955), and by Hess (1955). The modern concept for the Pacific Basin is that, from the sea surface down, the following crustal structure exists: (1) about 5 km of sea water, (2) about 0.5 km of sediment, and (3) about 5 km of crustal material assumed to be basalt from petrological and seismic information. At the bottom of the basalt layer the Mohorovičić discontinuity occurs at a level of marked density and seismic-velocity increase. Below the Mohorovičić discontinuity the material is thought (Hess, 1955), to be peridotite.

The strength of the crustal material is not well known, and much laboratory work in the field of high temperatures and pressures is necessary to determine the action of rock under stress in the earth's crust over long periods of time. The present ideas on the strength of the crust come from two sources: (1) laboratory experiments, and (2) deductions from seismic and geodetic information.

The compressive strength of basalt in the laboratory at 1 atmosphere and at room temperature varies from 460 to 4700 kg/cm^2 (Birch *et al.*, 1942). A recent paper by Robertson (1955) summarizes much of the data by Griggs, Bridgman, and others in this field. It is known that limestone under high temperatures and pressures flows plastically, but deformation of silicate rocks until recently was never accomplished in the laboratory (Griggs,

1954, p. 1258). Hydrostatic pressures cause limestone to become ductile, and the rupture strength of silicate rocks is raised. Original work by Robertson (1955, p. 1305), and compilation from the literature, indicates that at a mean pressure of 2500 kg/cm² (at a depth of about 10 km) the maximum shear stress of olivine diabase is about 2500 kg/cm². These are, of course, rupture-strength determinations, and it is to be expected that rocks subjected to stresses below the rupture strength will, in time, flow plastically; in other words the yield strength is less than the rupture strength and is time dependent.

From seismic and geodetic considerations Barrell (1915) and Benioff and Gutenberg (*in* Gutenberg, 1951, p. 390) estimate that the strength of the crust near the surface is about 1000 kg/cm². The strength below the surface has been estimated from the energy released by seismic ruptures to be, in the vicinity of deep-focus earthquakes at depths down to 700 km, on the order of 10 to 100 kg/cm² (Gutenberg 1951, p. 526); Bullen (1953) from the same data estimates 500 kg/cm². Benioff and Gutenberg (*in* Gutenberg, 1951) further estimate that, if isostasy is to be accepted, the order of magnitude of strength at a depth of 100 km is about 10 kg/cm².

All these estimates give general orders of magnitude, but the fact is that the behavior and strength of the crust below the surface are not well known, and there are no data which can be used in an engineering sense to calculate foundation strengths of the crust.

Isostasy can act only when, and if, the crustal material flows plastically at some depth in the subcrust. This depth is assumed from theoretical considerations to be less than 100 km (Benioff and Gutenberg *in* Gutenberg, 1951, p. 390). Down to depths of 700 km the earth's crust is capable of building up stresses, retaining them, shearing to relieve these stresses, and then keeping the ruptured surface open for periods on the order of a decade (Benioff, 1954). This, of course, implies that the material down to 700 km is a "solid" of considerable strength. The time intervals involved in isostatic adjustments and ideas on subcrustal convection currents, however, demand only an exceedingly slow plastic yielding so that such phenomena are not necessarily restricted to depths below 700 km. Scheidegger (1952) calculates that the speed of subcrustal convection currents can be on the order of "a few mm per year" which would still permit, because of a high viscosity, the building up of stresses. Although a very slow movement tending to relieve the stresses by plastic flow is assumed, the speed of relief of stress due to rupturing is so much faster that deep-focus earthquakes are possible in this "plastic" material. Thus, the strength of the crust down to 700 km, as indicated by deep-focus earthquakes, is considered compatible with the ideas of isostasy and subcrustal convection currents.

Formation of the Arch due to elastic deformation.—The crust of the earth is considered to be an elastic solid with definite elastic properties (Gutenberg *et al.*, 1951). If this is so then there probably would be elastic behavior of the crust under a load, or other stress. A number of papers have been written on this subject in which the crust has been considered as an elastic beam, an elastic sheet, or an elastic spherical shell. Gunn (1943, 1949), in studies of isobaric equilibrium, computed the elastic deformation to be expected under different types of load. In 1943 Gunn applied these concepts to the Hawaiian Ridge, and his computations and figures (1943, p. 383–390) clearly indicate how side deeps and bulges could form under the load of the Hawaiian Islands.

Vening Meinesz (*in* Poldervaart *et al.*, 1955) proposes that the great geosynclines and trenches have been formed by plastic buckling of the earth's crust, but that, outside this central plastic area, the crust behaves elastically. The formation of side bulges are indicated under this hypothesis (p. 324).

Gravity data by Vening Meinesz (1934, p. 106) across the Hawaiian Ridge at Oahu are best explained if the load of the Hawaiian Ridge is borne by the crust on a regional basis. Vening Meinesz states: "... The trough on the northeast might then be explained as the downwarping of the crust under this load, although it must be recognized that in this case another trough ought to be expected on the southwest side of the island ridge and nothing is found there." The marginal trough on the southwest side of the Hawaiian Ridge is not well developed, and along the profile of Meinesz's gravity stations it does not exist at all; it is nevertheless present on both sides of the ridge (Figs. 1, 4) and greater validity is thus shown for Vening Meinesz's original postulations.

Woollard (1951; 1954) made special studies of the gravity data available for Oahu including 26 new stations established by himself in 1948. He concludes that the best figure for density of the island mass is 2.3 gr/cc, that regional subsidence of the island to date has amounted to between 8000 and 10,600 feet,

and that it appears probable that all subsidence resulted from downward flexure of the crust (1951, p. 365). In a summary paper in 1954 on crustal structure beneath oceanic islands, Woollard concluded that subsidence through elastic yielding of the crust is a phenomenon to be expected with oceanic islands of volcanic origin. The present study, showing that the Hawaiian Ridge with marginal deeps lies within the great arches of the sea floor lends strong support to Vening Meinesz, Gunn, and Woollard in that it indicates flexural yielding of a probable elastic nature beneath the great load of the volcanic ridge.

The studies of both Gunn and Vening Meinesz indicate that the load of the Hawaiian Ridge is borne in part by the strength of the crust and in part by regional compensation spread out widely beyond the Ridge. The distance to regional compensation (crest of the Arch) shown by the profile of the Hawaiian structure (Fig. 7) is the approximate distance to compensation postulated by both writers.

A possible explanation for the features of the southern Hawaiian Ridge under an elastic bending hypothesis is illustrated in Figure 8. In Figure 8, *A*, the plastic deformation of the earth's crust (a fractured zone) has left a great linear welt on the sea bottom as described by Menard (1955). In *B*, volcanic rock fills the fractured areas and begins the build-up of the ridge to the surface where erosion starts to spread an apron around the base of the islands; some downbowing has already started. In *C*, downbowing has continued; faulting and volcanism are due to tension fractures in the newly formed elastic bulges on each side of the ridge. If the fractures created originally were at an angle to the surface, the many seamounts west of Hawaii, the filling of the southwest Deep, and the asymmetry of the two bulges are better explained. In *D*, downbowing, possible faulting on the west side, and partial filling of the west-side Deep give the present structural picture. The east-side Deep has been partially filled and is about as deep as the sea floor beyond the Arch. The raised position of the original welt allows some downbowing before the deeps are as low as the surrounding sea floor.

Bijlaard (1938, p. 15) believed that the crust would flow plastically before it buckled elastically, and Jeffreys (1952, p. 314) thought that fracture would occur before elastic "instability".

Whether the crust responds to a great load elastically or plastically certainly depends on rate of application and magnitude of the load, and on other crustal strength factors which are for the most part unknown. In considering the crustal response to load plastic as well as elastic hypotheses should be included.

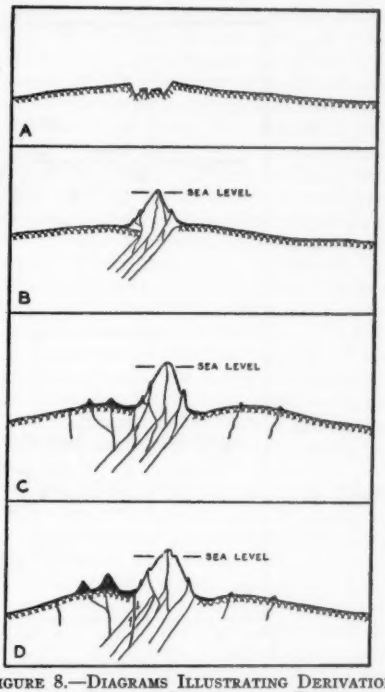

FIGURE 8.—DIAGRAMS ILLUSTRATING DERIVATION OF THE PRESENT STRUCTURE BY ELASTIC DOWNBOWING

(A) A great fracture zone welt in the ocean floor (after Menard, 1955); (B) Volcanic materials through fractures fill the central portions of the fractured zone and build a volcanic seamount to the surface; start of elastic downbowing and apron of erosional material; (C) Downbowing continues; tension fractures permit volcanism on the arches; (D) A combination of downbowing, volcanism, and faulting form the present structure.

Formation of the Arch due to foundation failure.—In contrast to the theory of formation of the Arch by downbowing of the elastic crust is the proposal that the presentt sructure is due at least in part, to the failure of the crustal foundation under the great load of the Hawaiian Ridge. This failure could have occurred, of course, before or during the elastic response previously discussed. This failure of the crust is largely a plastic failure, and thus the two hypotheses discussed in this study are elastic and plastic concepts.

In the Soil Mechanics' field of foundation studies the favored theory of foundation failures is that of Prandtl (Taylor, 1948, p. 572–579; Tschebotarioff, 1951, p. 221–224; Terzaghi and Peck, 1948, p. 167–169; Krynine, 1947, p. 207–209).

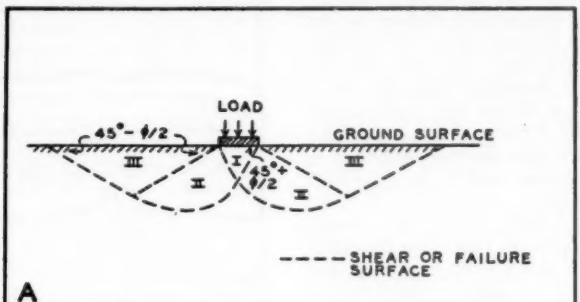

FIGURE 9.—DIAGRAMS ILLUSTRATING THEORIES OF FOUNDATION FAILURE

(A) Diagram illustrating Prandtl's plastic equilibrium theory of foundation failure; (B) Soil mechanics concepts of foundation failures with and without rotation of the load. (*After* Sowers and Sowers, 1951).

Prandtl's main concern was the plastic failure of metals under punches, but the theory has had wide influence in modern analyses of foundation failure problems. This theory, as applied to elongate, loaded areas, indicates three zones below such a structure whose foundation has failed (Fig. 9, *A*). Zone I moves down as a unit; Zone II is in a plastic state and pushes Zone III upward as a unit.

The shearing strength, *s*, of a cohesive material is equal to:

$$c + N \tan \phi$$

where c is the cohesion between grains, N the normal pressure on the plane of failure, and ϕ is the angle of internal friction, or the coefficient of friction. In material such as steel the shearing strength of the material is almost entirely in its cohesion and is due to intrinsic or internal pressure (Taylor, 1948, p. 319–321). The angle of internal friction is nearly zero for plastic clays and metals, and therefore the shearing strength of these materials is almost wholly in its intermolecular and intercrystalline cohesion. Where ϕ equals zero Prandtl's analysis of foundation failures reduces to:

$$P_{max} = 5.14c = 2.571q$$

where q is the compressive strength of the foundation material, and P_{max} is the maximum safe load or ultimate bearing capacity.

If the crust of the earth is thought of as an elastic continuum then failure, or more accurately, disintegration combined with plastic flow, occurs along trajectories of maximum shearing stress which are smooth lines without breaks: failure lines in an elastic continuum are nothing more than failure lines in an idealized fragmental mass, and the angle of internal friction, ϕ, is equal to zero (Krynine, 1947, p. 144). This failure results in side bulges of equal height as indicated in Figure 9*B*; but if the structure rotates, a common phenomenon in foundation failures of this kind (Tschebotarioff,

1951, p. 222), then the bulges are likely to be unequal, or not present on one side (Figure 9*B*, right). Where the foundation is not homogeneous or is weaker on either side then a rotation is likely to occur. Tschebotarioff (1951, p. 223–224) describes solutions of such structure rotations which are the same or very near the Prandtl solution. The side bulges, or arches, around the southern end of the Hawaiian Islands are wider than might be expected from a plastic failure of the described type: the wave length is more in accord with elastic response.

Whether the crust under the load of the Hawaiian Islands failed as a foundation depends on the strength of the crust under this load. The average load on the crust over the whole base of the Island of Hawaii is about 966 kg/cm² if the island is reduced to a symmetrical block which is about 2300 fathoms high and has a density of 2.3 gr/cc. The failure of the rock foundation, according to the Prandtl solution, would take place when the load exceeds about 2.57 times the compressive strength of the foundation material. If the range of surface basalt compressive strengths determined in the laboratory (460–4700 kg/cm²) are used, failure might occur if the load exceeded 1180 to a maximum of 12,100 kg/cm². The load computed above for the Island of Hawaii (966 kg/cm²) is about the same as the surface crystal strength of 1000 kg/cm² estimated by Gutenberg (1951, p. 390) and others from seismic and geodetic considerations. It can thus be seen that the loads available, even for this simplified picture, are of the magnitude to cause failure. The slow (geologically speaking) yielding of the rock due to flow under high pressures and temperatures and stresses below the rupture strength of the rock, especially in weakened sectors, could result in the failure of the foundation under surface loads less than necessary for shorter-term rupture or yielding.

Tschebotarioff (1951, p. 222) notes that most foundation failures are of a rotational nature so that the structure falls to one side because of inequalities of strength in the foundation materials. Cracking and fissuring of the earth's crust on a large scale were necessary to permit the effusion of volcanic materials to form the Hawaiian Ridge; the implication is that the crust in this sector has been weakened by high temperatures, by actual cracking, and by the presence of fissures filled with liquid magma. It seems logical, therefore, to assume that the foundations of the Ridge are weaker than is the crust elsewhere, and that these weaknesses are not necessarily spread equally beneath the load. If the rock foundation of the Hawaiian Ridge failed in a manner similar to that envisaged by Prandtl for soil foundations, it seems likely that there was rotation of the load which caused the side bulges to be unequal. Such a rotation caused by failure of the foundations under the southern end of the Hawaiian Ridge (falling off to the east) might explain the marked difference in elevation of the side bulges and the unequal depths of the deeps on the northeast and southwest sides of the Ridge. The marked bulging upward of the crust on the southwest side of the Ridge would cause tension cracks to form and open the way to more volcanism along these fissures than would be the case on the northeast Arch. These new volcanic vents would, if sufficiently active, create new seamounts or marked irregularities on the Arch, as is the case to the west of the Island of Hawaii. The presence of the Arch on the northeast and southwest sides indicates that this rotation did not take place until after the failure was well underway. The failure did not begin until the islands had reached the surface and erosion had started to spread an apron of sediments around the island base.

With a Prandtl type of foundation failure fissures would be created along the bases of the island structures and result in the pouring out of lavas to help fill the marginal deeps created along the Ridge by the sinking of the structure and the formation of the bulges. Menard (1956) believes (from other considerations) that partial filling of marginal deeps by submarine volcanic material is necessary to obtain the volume, including normal sedimentary and volcanic materials, required to fill marginal deeps and establish the profile of equilibrium seen on the aprons sloping away from older eroded islands and former islands.

Combination of elastic and plastic yielding.—The writer holds no great brief for a Prandtl type of foundation failure to explain the present structure of the southern Hawaiian Ridge. It is an interesting possibility which should certainly be considered in this type of problem. An elastic yielding of the crust, or a combination of the two types of yielding, is possibly a better explanation.

A sequence of events in the slow build-up of a great seamount or ridge on the ocean floor is possibly as follows:

(1) Fractures or fissures open up in the earth's crust through which magma can rise to the surface to form the ridges and volcanoes.

(2) The first effects of the rising volcanoes are to squeeze to the sides the unconsolidated sediments on the sea floor.

(3) As the volcanoes build higher the earth's

crust is, at first, entirely competent to hold the structure without plastic or elastic response.

(4) At a certain point in loading, downbowing of the crust occurs, side bulges and marginal deeps are formed as an elastic response; plastic flow to the sides occurs beneath the crust. This is the possible state of great seamounts and moats unfilled by lavas or sediments.

(5) As the loading continues (or a heavy load causes yielding over a long time) a point is reached when the ultimate bearing capacity of the foundation is exceeded, and the foundation fails in plastic shear. This possibly occurs only on those seamounts large enough to be near the surface or actually become islands and may not have occurred on all islands.

(6) Regional compensation of the load is finally achieved.

If an island foundation fails plastically the marginal deeps will tend to fill with magmatic material extruded from the fractures around the base of the structure. Seismic studies and a recent study by Menard (1956) indicate that normal volcanism and sediment from islands are insufficient in volume to fill marginal deeps to form the smooth aprons found around many ancient sunken islands or present islands. The added volcanic material from the fault planes around the base of a structure whose foundation has failed might have supplied this necessary additional material.

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UPPER MIDDLE ORDOVICIAN STRATIGRAPHY OF FILLMORE COUNTY, MINNESOTA

BY MALCOLM P. WEISS

ABSTRACT

Strata between the Glenwood and Maquoketa formations in Fillmore County, Minnesota, and adjacent areas of Minnesota and Iowa, are representative of these same beds cropping out farther to the north and west in the Mississippi Valley. They have received little modern study, and have heretofore not been subdivided on lithologic grounds alone.

The Platteville Formation is a dominantly carbonate sequence between the Glenwood Formation and the shale of the Decorah Formation; it consists from the base upward of the sandy dolomitic limestone of the Pecatonica Member, with multiple corrosion zones at its top; the limestone of the McGregor Member, with a corrosion zone near its base; and the Carimona Member, of interbedded limestone and minor shale with a conspicuous bentonite at or near its base in Fillmore County. The Decorah shale includes the Spechts Ferry bentonite near its base, contains hematite oölites at the top toward the west, and is not divisible into the members recognized to the southeast. The Galena Formation consists of three members: the Cummingsville cyclical limestone and shaly limestone at the base; the Prosser limestone, which has a group of corrosion zones at its top; and the Stewartville dolomitic limestone, which has a feldspathized shale near the base. The Dubuque Formation, of interbedded limestone and shale, contains two feldspathized shale beds and is thought to be the highest Middle Ordovician rock in the area.

Large collections of macrofossils were made, and their stratigraphic distribution recorded in detail and summarized in a faunal list. These data show clearly that the terms Black River and Trenton cannot be applied meaningfully to Minnesota Ordovician rocks on the basis of macrofossils.

The geologic history recorded by these rocks and fossils is interpreted as marine transgression continued from the time of deposition of the St. Peter-Glenwood, stillstand and then regression during Decorah deposition, continued regression and stillstand during Galena deposition, and renewed transgression of greater magnitude from Dubuque through Maquoketa deposition. The historical implications of corrosion zones are discussed.

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Introduction and Acknowledgments

The classification and understanding of the nature and distribution of upper Middle Ordovician rocks in southeastern Minnesota have evolved during a period of more than 50 years. Contributuors to the solutions of these problems have based their work on different concepts and have worked with different geographical emphasis. For example, the papers by Winchell and associates (1884–1897) and Sardeson (1892–1897) produced a rock nomenclature that is based partly on faunal zones. In a series of papers published from 1929 to 1940, G. M. Kay contributed greatly to the knowledge of the Ordovician of the upper Mississippi Valley, but much of his work is not directly applicable to Minnesota rocks because of facies changes. The rock classification of Stauffer and Thiel (1941) is an adaptation of Kay's names to units distinguished in the older Minnesota reports. This paper classifies these rocks on lithic characters alone (Fig. 1) and shows the distribution of the faunal elements in the rock (formational) units.

With the exception of recent unpublished work, Ordovician paleontology in Minnesota has been characterized by vague stratigraphic data. The conclusions and descriptions contained in this paper result from the study of more than 60 measured sections and wells in and around Fillmore County.[1]

This report is an outgrowth of a doctoral dissertation prepared for the University of Minnesota and abstracted in Weiss (1953). Aid, encouragement, and criticism by the members of the Department of Geology and Mineralogy is gratefully acknowledged, particularly that of W. C. Bell (now of the University of Texas), who directed the work. The Minnesota Geological Survey supported the field work and some of the drafting and laboratory work. A. F. Agnew kindly made his dissertation available to the writer and joined in several field conferences. Thanks also go to R. L. Bates, W. C. Bell, and E. C. Dapples, who criticized this paper.

[1] Records on file at the Minnesota Geological Survey.

Stratigraphy

Controlling Principles

Problems of sequence, distribution, and changes of rock strata can be resolved only if the distinction between lithostratigraphy and biostratigraphy is kept clearly in mind. Formations and members are rock units; they should carry no connotation of time nor be defined by faunal content. In this work all rock units were identified or defined, and their limits described, before fossil collections were studied.

Different procedures would obviously be necessary in rocks that are totally or nearly unfossiliferous, and in rocks that are homogeneous but contained abundant fossils. For the former, "lithostratigraphy" is the only key; and for the latter, changes in the fauna ("biostratigraphy") might provide the only means of subdividing the rock mass. Most sedimentary rocks are somewhere intermediate to these "pure" conditions; the rocks with which this paper is concerned have numerous changes of facies. Rocks such as these must be described and classified according to their physical qualities, and faunal data must be related to this framework. Such practice affords a classification useful to everyone, whether paleontologist or not, and makes possible the revision of interregional time correlations without disturbing the local rock classifications. Furthermore, it has the realistic effect of relating fossil faunas to the rock units containing them in the same way that the living faunas were related to (in fact largely controlled by) the environments of sedimentation. There are fossils which were unrelated to the bottom environment in life, but these are a minor part of most faunas, including those of the area under discussion.

A sequence of lithostratigraphic and biostratigraphic units is interpreted as a succession of environments; each replaces those earlier both horizontally and vertically. Thus, uncon-

Sardeson, 1892, 1897 (some units too thin)	Winchell & Ulrich, 1895, 1897	Kay, 1935a, 1935b	Agnew, et al., 1956 (Quimbys Mill absent)	Stauffer & Thiel, 1941	This Report
Beds 11 to 14	Richmond limestone	Maquoketa Formation	Maquoketa Formation	Maquoketa Formation	Maquoketa Formation
10 Triplecia Bed (Oxoplecia)	Utica shales	Dubuque Formation	Dubuque	Dubuque Member	Dubuque Formation
9 Maclurea Bed (Maclurites)	Trenton: Limestone: Maclurea Beds	Stewartville Formation	Non-cherty	Stewartville Member	Galena Formation: Stewartville Member
8 Lingulasma Bed	Trenton: Limestone: Galena limestone	Prosser Formation	Cherty	Prosser Member	Prosser Member
7 Camarella Bed (Parastrophina)	Trenton: Shales: Upper Galena shale	Galena Group	Galena Formation	Galena Formation	Galena Formation: Cummingsville Member
6 Orthisina Bed (Vellamo)	Trenton: Shales: Basal beds				
5 Fucoid Bed	Black River: Shales: Upper 3rd	Decorah Formation: (Ion Member)	Decorah Formation: (Ion Member)	Decorah Member: (Ion Submember)	Decorah Formation
4 Stictopora Bed (Rhinidictya)	Black River: Shales: Middle 3rd	(Guttenburg Member) S. F. Member	(Guttenburg Member) Sp. F. Member	(Guttenburg Submember) Spechts Ferry Member	
3 Stictoporella Bed	Stones River: Shales: Lower 3rd	Platteville Formation: McGregor Member	Platteville Formation: McGregor Member	Platteville Formation: McGregor Member	Platteville Formation: Carimona Member / McGregor Member
2 Bellerophon Bed	Shaly limestone	Pecatonica Member	Pecatonica Member		Pecatonica Member
1 Buff Limestone Bed	Building layers	Glenwood Member	Glenwood Member	Glenwood Member	Glenwood Formation
St. Peter Sandstone	St. Peter Sandstone	St. Peter Sandstone	St. Peter Sandstone	St. Peter Sandstone	St. Peter Sandstone

FIGURE 1.—REPRESENTATIVE CLASSIFICATIONS OF MIDDLE ORDOVICIAN ROCKS OF THE UPPER MISSISSIPPI VALLEY

Applied to Fillmore County, Minnesota, and compared with the usage of this report. The sum of the typical thicknesses cited by Sardeson for his "beds" is less than the physical thickness of these rocks measured by the writer. Equivalent modern names for Sardeson's "beds" are given in parentheses.

formities are here accepted only upon physical evidence. A number of diastemic surfaces occur in the Ordovician limestone strata of Minnesota (Weiss, 1954b) and elsewhere in the upper Mississippi Valley. Whether any of these surfaces can be shown to have regional extent or significance remains to be seen. The Platteville-Decorah disconformity in south-central Wisconsin, described by Agnew (1948, Ph.D. thesis, Stanford Univ.; 1956), is not present in Minnesota.

Glenwood Formation

The sequence of sandy and shaly beds that overlies and is superficially distinct from the St. Peter Sandstone in outcrop ranges in thickness from 4 to 5 feet to twice that much. North and west of Preston, Minnesota (Pl. 1), beds of coarse sandstone cemented with calcite form the major part of the interval. Use of "Glenwood Formation" here is provisional and arbitrary.

The upper limit of the Glenwood is sharp and can be precisely defined from field evidence as the level of change from sandstone, shale, or sandy shale to limestone or dolomite or some mixture thereof. Scattered grains of quartz sand occur in the basal carbonate beds of some localities. The upper part of the Glenwood commonly is cemented to some extent by secondary calcite but need not be confused with the limestone or dolomite above; it is clearly a fabric of closely packed quartz grains and interstitial calcite cement, rather than a carbonate matrix with floating quartz grains. A shaly interval less than 6 inches thick is present at the top of the Glenwood at most localities (both the uppermost sandstone and the lowermost carbonate rock are argillaceous), and the contact is placed at the change from shaly sand to shaly limestone or dolomite. At some localities the uppermost Gelnwood is almost entirely friable sand, and carbonate rock lies directly on it.

Platteville Formation

General.—The Platteville Formation is the unit of limestone and dolomitic limestone between sandy shale (Glenwood) at the top of the St. Peter Sandstone and the green calcareous shale of the Decorah Formation. It was named by Bain (1905, p. 19) for exposures near Platteville, Wisconsin. The formation has been variously interpreted in the type area, but the top has recently been precisely defined by Agnew and Heyl (1946). The upper member (Quimbys Mill) of the Platteville Formation of the type area and the lower member (Spechts Ferry) of the Decorah Formation are not present in Fillmore County, nor elsewhere in Minnesota.

The name Beloit Formation was proposed by Sardeson (1896, p. 360) for the "Trenton" beds of Wisconsin below the Galena Formation. It was composed of "beds" 1–5 in Sardeson's classification and is equivalent to what is today known as the Platteville and Decorah formations. Although Sardeson did not describe a typical measured section nor specify the characteristics of the boundaries (other than faunal), his meaning was precise because of the extensive synonymy cited under the name Beloit. Surprisingly, Sardeson's proposal seems to have been ignored completely by everyone except Agnew (1948, Ph.D. thesis, Stanford Univ.; Agnew *et al.*, 1956). Beloit Formation is a valid name with considerable priority over both the names Platteville and Decorah, and it is unfortunate that Bain (1905) and Calvin (1906) gave it no notice when they proposed their names. The name Beloit has no currency in the literature except in the works of Sardeson himself, and no good can be accomplished by reviving it at this time.

In southeasternmost Minnesota the Platteville Formation is composed of three members: in ascending order, the Pecatonica, McGregor, and Carimona. The Pecatonica Member is present only in the southeastern part of Fillmore County and thins rapidly northwestward. The McGregor is present throughout the county, is irregular in thickness, and lies directly on the Glenwood Formation where the Pecatonica Member is absent. The bottom and top contacts of the Platteville Formation are gradational within rather narrow limits. There is a progressive change in the fauna from the bottom to the top, but no evidence for what might be called a profound change of fauna.

Pecatonica Member.—The "Pecatonica dolomite," a moderately thick-bedded buff or light-brown dolomite, argillaceous and nearly free from fossils, gradational into limestone above and St. Peter Sandstone below, was named by Hershey (1894, p. 175) from exposures in the Pecatonica valley north of the Illinois-Wisconsin line. He clearly included what is now Glenwood in his Pecatonica, but Kay (1935a) restricted the name to the carbonate unit. In Fillmore County (Pl. 1) the

Pecatonica is a magnesian or dolomitic limestone with floating quartz grains concentrated near the top and toward the base. The rock is light brownish gray, weathers to pale orange, and is generally fine-grained, medium- to thick-bedded, microvuggy, and burrowed. The sand is concentrated in the burrow fillings, is poorly sorted, and is composed for the most part of the well-rounded and frosted grains typical of the St. Peter and Glenwood formations. Except for the uppermost few inches the member is unfossiliferous. The Pecatonica is distinguished from the McGregor Member by its coarser grain size, thicker bedding, smooth bedding planes, and generally dolomitic and sandy nature. The two units, however, grade into each other. The placement of the member boundary at the top of the group of three corrosion zones (Weiss, 1954b) is therefore somewhat arbitrary.

There are at least two other corrosion zones[2] in the Pecatonica in addition to those at the top of the member. From the top of the Pecatonica down, the corrosion zones or corroded surfaces themselves differ as follows: the uppermost is very pyritic and has small relief; the second is not conspicuous, has little pyrite or collophane, is not much pitted, and can be distinguished only on a fresh face; the third is very conspicuous, thick and irregular, and associated with much sand, pyrite, and collophane; the fourth and fifth resemble the third but have less relief and less collophane and sand.

Acid-insoluble materials (quartz sand, clay, silt, and pyrite) make up 5–20 per cent of the rock and average 12 per cent. Collophane occurs as shiny black pebbles and granules and as reddish-brown flakes in the burrow fillings. The rock also contains some conodonts, scolecodonts, and scaphopods (?). The clastic calcite mosaic shows numerous fossil fragments and about 15 per cent of dolomite in thin section. Most of the pyrite is associated with organic debris. Very fine quartz grains, both rounded and splintered, show in thin section, in addition to the coarser grains seen in hand specimens. All quartz grains show some etching and replacement by calcite along grain boundaries.

The Pecatonica Member ranges in thickness from about 7 feet east of Mabel to a feather edge about 20 miles northwest. It thickens as it continues into Iowa, Wisconsin, and Illinois. Majewske (1953, M.A. thesis, Univ. Minnesota) reported it to be about 1 foot thick at Minneapolis, to thicken eastward, and to have a corrosion zone near the top. W. C. Bell (Personal communication) recognized about a foot of Pecatonica in the vicinity of Rochester, Minnesota, and found corrosion zones at the top and within the member there.

Correlation of corrosion zones from Fillmore County to Minneapolis or to Illinois (Templeton and Willman, 1952) is unwarranted at present but may some day be possible. Bell has found them in the Pecatonica in southwestern Wisconsin, and Prokopovich (1955, Fig. 2) reported one from the subsurface of Iowa. It does appear, however, that the corroded surface at the top of the member in Fillmore County may be the equivalent of that at the top in the Rochester area. East and northeast of Chatfield, in Olmsted County (Bell, personal communication), as much as 2 feet of atypically argillaceous and dolomitic limestone at the base of the McGregor contains two or more corrosion zones and may represent the Pecatonica.

McGregor Member.—The McGregor Member was named by Kay (1935a, p. 286) from an exposure near McGregor, Iowa, and defined as the limestone unit succeeding the Pecatonica and underlying the Spechts Ferry Member of Kay (1929). In Fillmore County the McGregor is a unit of limestone that is light olive gray to yellowish gray, generally microgranular, thin and crinkly bedded, hard and brittle, and fractures conchoidally. Some conspicuous shaly bedding planes divide the unit into ledges, and it weathers rapidly to a yellowish-gray slabby outcrop. A corrosion zone with little collophane and no quartz sand is present 4 feet above the base of the member where it is underlain by the Pecatonica. Where the McGregor lies directly on the Glenwood Formation the corrosion zone approaches the base as the member thins northwestward. Near the top of the unit are two intervals, separated by 6 inches of hard limestone, which are much more shaly than the rest of the member. Each interval is about a foot thick and weathers to a conspicuous recess in the outcrop. The topmost 1–2 feet of the member has some of the brittle, splintery character of the "glassy McGregor" at the top of the member in the Wisconsin-Illinois mining district. This character is best developed in the southeastern part of the Fillmore area. The

[2] Corrosion zones are rough to dove-tailed surfaces of small relief with concentrations of pyrite and commonly also collophane and quartz sand in the rock over the corrosion surface. They are believed to be surfaces of submarine solution and of accumulation of insoluble debris.

McGregor is sparsely fossiliferous but contains much organic debris. Most fossils seem to be inlaid into the top surfaces of the rippled beds; their convex surfaces are uppermost. Weathering along the shaly partings produces slabs of limestone having fossils clearly exposed in relief.

The McGregor is distinguished from the Pecatonica by being yellowish, olive, or bluish gray rather than brown or brownish gray, by the thin, rippled bedding, and by the fact that it contains little dolomite. The basal 2 inches of the McGregor is dolomitic and very sandy where it overlies the Pecatonica. Where the McGregor lies on the Glenwood the basal beds are characteristically salted with floating quartz grains, but in the Fountain-Preston area no sand could be found in the McGregor. The McGregor apparently is conformable[3] with the Glenwood; a thin shaly interval commonly bridges the sand-to-carbonate transition. It is distinguished from the overlying Carimona Member by its finer grain size, gray color (instead of brown), thinner and more wrinkled bedding, and the conchoidal fracture. In some respects the two members are transitional. The upper 2 feet of the McGregor is thin- to medium-bedded, and the beds are noticeably less wrinkled than those below. This upper part is also of mixed microgranular and very fine-grained rock, of which the former is more conspicuous in the McGregor and the latter is typical of the Carimona.

Acid-insoluble materials (olive-gray clay, pyrite, phosphatic organic fragments, conodonts, and scaphopods (?)) make up 2–30 per cent of the rock and average 5–10 per cent of it. Quartz sand commonly occurs in small amounts in the basal layers. The clastic calcite matrix shows many fossil fragments coated with pyrite cubes in thin section. Tiny rounded grains of collophane also occur. The scattered quartz grains are corroded and partly replaced by calcite. The small clay fraction, except on parting planes, is thinly dispersed through the matrix. Only one thin section, from a bed adjacent to the Pecatonica, shows as much as 7 per cent of dolomite. These results can be reconciled with those of Stauffer and Thiel (1933, 1941) only because all but one of their samples came either from the Pecatonica Member or from the dolomitic Platteville rock of the Twin Cities. Irregular yellow-weathering patches that occur in the McGregor probably are dolomitic, and individual grab samples may run higher in dolomite than adjacent ones.

The McGregor Member is 20 feet thick in the southeast, where it overlies the Pecatonica Member, and thins to about 16 feet in the central part of the county. The thinning is irregular, and anomalous thicknesses greater than at near-by outcrops occur at a few places in the central and northern parts of the county. The southeastward thickening of the member continues into southwestern Wisconsin, where it is about 25 feet thick. From one to several corrosion zones have been found in favorable exposures between Minnesota and Illinois (W. C. Bell, personal communication), but possible regional relationships among these are not as yet understood.

The McGregor thins northwestward to 14 feet or less in the vicinity of Rochester, Minnesota, and northwest of there gradually changes character until the name "McGregor" is no longer applicable in the vicinity of Cannon Falls. In the Twin City area the equivalents of the McGregor total about 26 feet in thickness. At Cannon Falls the Platteville consists wholly of the Carimona Member and the Hidden Falls Member (Sloan, 1956), first defined by Majewske (1953, M.A. thesis, Univ. Minnesota; 1956; *see* also Weiss and Bell, 1956, p. 60–61). Thus the depositional environment of the typical McGregor, or of the Magnolia and Mifflin, must not have been continuous from the Twin Cities southward but probably was continuous farther east in Wisconsin.

Carimona Member.—The name Carimona Member was proposed by Weiss (1955) for the few feet of limestone and interbedded limestone and shale at the top of the formation and underlying the freen shale of the Decorah Formation. The limestone beds of the Carimona Member are pale reddish brown, fine-grained and very fine-grained, medium-bedded, and have smooth to broadly rippled surfaces. They are separated by beds of yellowish-brown to olive calcareous fissile poorly fossiliferous shale that increase in thickness from bottom to top of the unit. Near the base of the member a thin unfossiliferous chocolate-colored shale is overlain by a conspicuous bed of bentonite 2–4 inches thick. Locally there are thin patches of shale over the bentonite.

[3] If conformable, then the corrosion zones in the Pecatonica must be either (1) diastems wholly within the Pecatonica-McGregor carbonate facies, or (2) diastems that cross a facies boundary from the Pecatonica into the Glenwood. In the absence of good evidence, the author believes the latter possibility more probable (Pl. 5).

Carimona limestone beds have a hackly fracture normal to the plane of bedding that contrasts with the smooth, conchoidal fracture of the underlying McGregor beds. The two uppermost beds of limestone are somewhat less brownish and more gray than the rock below, are mottled with patches of medium gray, and are pyritic. The Carimona Member is moderately fossiliferous.

The Carimona is distinct from the McGregor in grain size, color, fracture, bedding, and shale distribution, although the uppermost McGregor shows some qualities of Carimona bedding and grain size. The contact is placed below a thin yellowish-brown or chocolate-colored calcareous unfossiliferous fucoidal shale. In and northwest of the type area beds of limestone intervene between this shale and the bentonite (Pl. 2, fig. 1), but in the southeast the shale and bentonite are separated by less than an inch of limestone. (*See* also Bell, 1954.) The Carimona Member extends upward to include the highest bed of Carimona-type limestone and consequently includes a few thin beds of greenish shale of Decorah type. Although the Platteville and Decorah formations are clearly transitional and conformable, the arbitrary contact between them is evident at every exposure.

Carimona limestone typically contains 5–10 per cent of insolubles, generally clay, less commonly silt. Pyrite occurs in most residues but is more common in the upper part. Conversely, phosphatic materials (mostly conodonts, and some fragments of scaphopods (?)) are more abundant in the lower part. Thin sections show the limestone to be a calcarenite with abundant fossil fragments coated with pyrite and scattered particles of collophane.

The member thickens westward across Fillmore County from 3 to 6 feet and extends to Cannon Falls and the Twin Cities (3 feet, 9 inches). It thins southeastward into Iowa and at McGregor is only a few inches thick, including the bentonitic limy shale that represents the Carimona bentonite. In this area, where the Spechts Ferry is a recognizable rock unit, the Carimona directly underlies it. Agnew *et al.* (1956, p. 265) state:

> "The western edge of the Quimbys Mill is an approximate north-south line about 5 miles west of the east line of Grant County, Wis. (fig. 35). Recognizable beds of the unnamed limestone unit [Carimona Member] are seen as far east as the mouth of the Wisconsin River, which is 30 miles west of the Quimby Mills pinchout. Between the mouth of the Wisconsin River and the central part of Grant County there are beds of indeterminate assignment, although presently they are recognized as having some of the characteristics of the unnamed limestone member."

Bell (Personal communication), who has seen many of the outcrops referred to by Agnew, concurs.

Although the detailed facies relations in this area are not yet worked out, it is certain that the Carimona and Quimbys Mill members are lithologically distinct, and that they occupy the same stratigraphic interval northwest and southeast of the area in question.

A conspicuous bentonite seam lies wholly within or underneath the Carimona Member and about 4 feet 6 inches (in Fillmore County) below a less conspicuous bentonite in shale at the base of the Decorah Formation. These bentonite beds converge southeastward and diverge northwestward (Bell, 1954). The bed in the Carimona Member is that one in Minnesota that has been mistakenly correlated with the Spechts Ferry bentonite in Iowa (Bell, 1954; Weiss, 1953, Ph.D. thesis, Univ. Minnesota). These bentonites are oblique to the boundaries of the several rock units, after the manner of time surfaces, so that within Fillmore County alone the thickness of Carimona rock under the bentonite decreases southeastward from more than 1 foot to less than 1 inch. At McGregor, Iowa, the thin Carimona rock lies between the two bentonites. Figure 11 of Weiss and Bell (1956) shows these relations but is in error in recording the upper bentonite at Cannon Falls, Minnesota.

Decorah Formation

The unit of green calcareous shale between the Platteville and Galena limestone formations was named the "Decorah (Green) shale" by Calvin (1906, p. 60, 84–87) from exposures in the town of Decorah, Iowa.

Its members (Kay, 1929; 1930; 1935a) are recognizable lithic units at their type sections and eastward. Agnew has recently (1948, Ph.D. thesis, Stanford Univ.; 1956) been successful in tracing them into and east of the mineral district and found that each of the members thins eastward. Kay did not attempt to distinguish the members northwest of Decorah. Where the formation grades to shale, between the type sections of the members along the Mississippi and near Decorah, he identified the top of the Spechts Ferry by a hard bluish limestone, in which Agnew (1948, Ph.D. thesis, Stanford

Univ.) found phosphatic pellets and steinkerns; the top of the Guttenburg by a concentration of *Glyptorthis*; and the top of the Ion by a concentration of *Prasopora* (Kay, 1929, 1932; Kay and Atwater, 1935).

Stauffer and Thiel (1941) accepted the member names of Kay and correlated them precisely (following Kay, 1940) with certain of Sardeson's (1892) faunal zones. However, in making the Spechts Ferry equal to the *Stictoporella* bed, they, like Kay, included some limestone (Carimona) at the base in Minnesota, whereas Kay's Spechts Ferry in northeastern Iowa is a shale unit. Stauffer and Thiel ascribed no distinctive lithologic features to the members in Minnesota except a pebbly corrosion zone at the top of the Spechts Ferry and a bed of "brassy" oölites (see below) at the top of the Ion. The former is not present in southeastern Minnesota, and the latter occurs only north and west of central Fillmore County.

In this report the unit of green calcareous shale (Pl. 2, fig. 2) between the limestone of the Platteville Formation and the limestone of the Galena Formation is called Decorah, without differentiation into members. Therefore the Decorah Formation of this report includes the shaly upper part of the Spechts Ferry, the Guttenburg, and in places the Ion of Stauffer and Thiel (1941).

In Fillmore County and adjacent areas of exposure the Decorah Formation is a shale, grayish yellow green to greenish gray, calcareous, generally fissile, but more massive and blocky where coquinoid layers are absent or sparse. It contains numerous coquinoid limestone layers concentrated near the base and top, and subordinately near the middle. There is a 1- to 2-inch bentonite near the base, under the lowest substantial shell bed. The bentonite is about 3 feet above the base of the formation in the west and 16–18 inches above it in the southeast. This clay seam is the "putty layer" of the Twin City section, is continuous with the Spechts Ferry bentonite, and was called the Hounsfield bentonite in Iowa by Kay (1931) (see Weiss, 1953, Ph.D. thesis, Univ. Minnesota; Bell, 1954). The Decorah is the most fossiliferous unit in the Minnesota Ordovician. The fauna is dominated conspicuously by brachiopods and bryozoans, and most of the fossils occur in the coquinites or coquinoid limy layers.

The base of the formation is evident where the shale lies on the uppermost hard limestone of the Carimona Member (Pl. 2, fig. 1). The top of the formation displays two aspects. As far west as the center of the county the basal nodular beds of Galena limestone lie directly on green shale. Farther north and west the uppermost shale is replaced by shale and limestone heavily charged with "brassy" oölites. There the interval of nodular beds at the base of the Galena is thinner than where the "brassy" oölite beds are absent. Thus the "brassy" oölite facies at the top of the Decorah displaces clean green shale west of the middle of Fillmore County. Regardless of which facies is present there is everywhere a concentration of the massive bryozoan, *Prasopora*, in the transitional beds of shale and nodular limestone in the base of the Galena Formation (Weiss and Bell, 1956, Fig. 12). The horizon of *Prasopora* is probably that which marks the top of Kay's Ion Member in Iowa.

PLATE 2.—PLATTEVILLE AND DECORAH EXPOSURES

FIGURE 1.—Fountain North section (F96). The Carimona Member extends from the hammer head to the top of the highest ledge. The Carimona bentonite lies in the slot behind the end of the handle. This outcrop is typical of the Carimona Member. (NE¼ sec. 27, T. 104 N., R. 11 W.)

FIGURE 2.—Sugar Creek section (F147). Thick blocky shale of the middle Decorah Formation. Coquinoid layer above hammer head is 19 feet above the base of the formation. (N line of NW¼ sec. 9, T. 103 N., R. 11 W.)

PLATE 3.—STEREOGRAPHIC PHOTO-MAP OF THE MAHOOD'S CREEK AREA WEST OF WYKOFF, MINNESOTA (T. 103 N., R. 12 W.)

Confluence of Mahood's Creek from the south and Spring Valley Creek from the west marked by a circle. The several exposures of the Mahood's Creek measured section identified by letters. The top of the Carter's Creek section (F93), Prosser's Corner, and the section west of Prosser's Corner (F165) (Stauffer and Thiel, 1941) are also marked. The limits of Prosser's ravine, at the southeast, marked by quotation marks. Air photos VZ-2D-42 and 43, U. S. Department of Agriculture, Production and Marketing Administration.

FIGURE 1

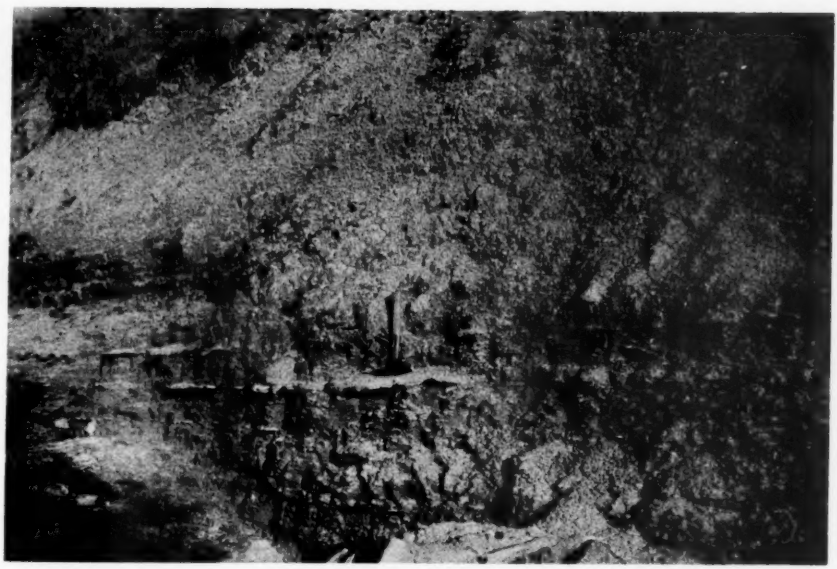
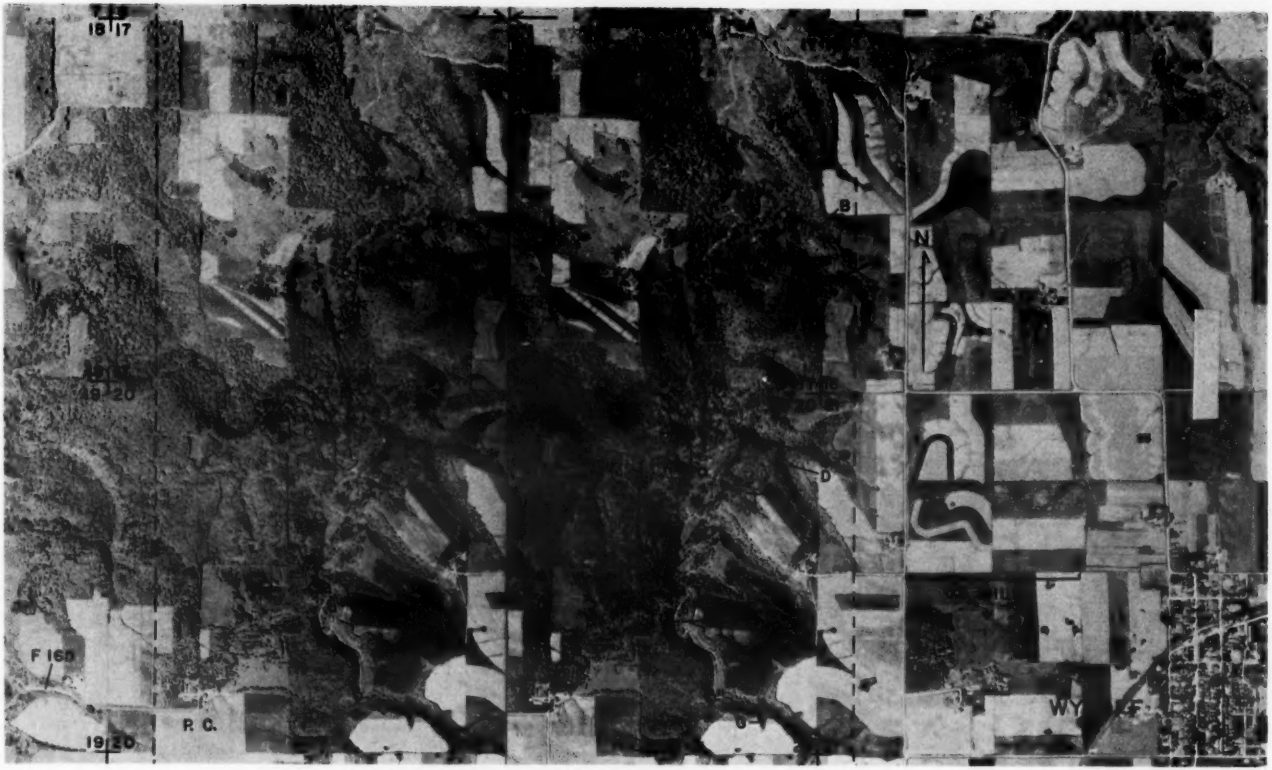


FIGURE 2

PLATTEVILLE AND DECORAH EXPOSURES



STEREOGRAPHIC PHOTO-MAP OF THE MAHOOD'S CREEK AREA
WEST OF WYKOFF, MINNESOTA

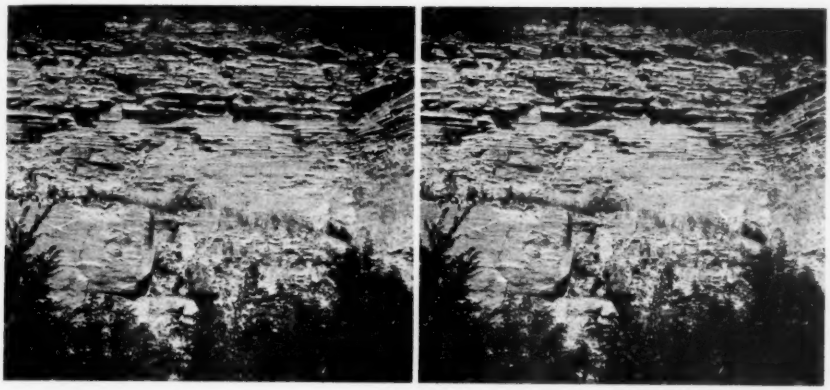


FIGURE 1

FIGURE 2

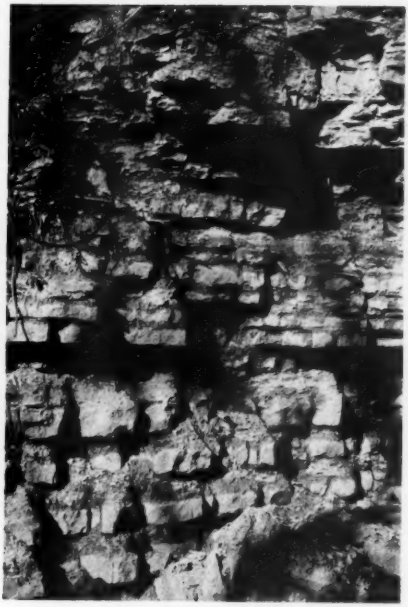


FIGURE 3

GALENA AND DUBUQUE EXPOSURES

STEREOGRAPHIC PHOTO-MAP OF THE MAHOOD'S CREEK AREA WEST OF WYKOFF, MINNESOTA

The petrography of Decorah shale at the Twin Cities has been described in detail by Thomes (1937, M.A. thesis, Univ. Minnesota), and other data are available in Grout and Soper (1919). Thomes found considerable potassium feldspar throughout the shale and some even in the limy layers. The lmy layersi in the Decorah Formation are of two types, Near the base, particularly in the southeast- are several layers of soft punky light-greenish. gray limy marlstone, usually containing very few fossils. Most of the remainder of the limy layers are hard coquinites or coquinoid limestones. Some of these contain considerable green shale in streaks and pockets, but most are nearly free from it. Most of the fossils in the formation are contained in these coquinites, although there is a scattering of fossils through much of the shale, and microfossils occur abundantly in the shale. The possible significance of these layers is discussed under geologic history.

Thomes described (1937, M.A. thesis, Univ. Minnesota) "brassy" oölites only briefly as limonite oölites formed about a core of mineral grain or fossil fragment. The writer found that the "brassy" oölites from Fillmore County are predominantly of hematite. Most have an outer crust of goethite, and a few are wholly goethite. They are the color of polished bronze, and the goethite forms a hard shiny outer coat. All are true oölites; most are solid, but a few have geodal centers. Many have cores or nuclei of calcite or hematite fossil fragments. Although the cores are of a variety of shapes the successive layers of each oölite compensate for the irregular cores and produce a nearly uniform flattened ellipsoidal form. They are almost uniform in size, about 1 mm in maximum diameter. These particles are clearly related to the "ironshot" described by Pettijohn (1949, p. 341), although not precisely similar to those he described.

As the "brassy" oölites weather out of shale or shaly limestone they are clean and bright. A thin section of fresh oölitic limestone from the upper Decorah at Carter's Creek (Pl. 3, *F*93) shows that the oölites are coated almost completely with fine cubes and needles of pyrite. The pyrite is probably a secondary product developed by the reduction of some of the iron in the oölites by organic matter entrapped in the beds.

At about the middle of the formation at Carter's Creek there is a bed of hematitic fossil debris. Fragments have been replaced by hematite and some goethite, and they present exactly the same surface appearance as the true oölites. The average size of these particles is much greater than that of the oölites. This horizon appears to be represented at Sugar Creek (Pl. 2, fig. 2) by a very thin streak of well-weathered fragments or oölites. Agnew (1948, Ph.D. thesis, Stanford Univ.) reported "discoidal, hematitic and limonitic concretions" from the Guttenburg Member in wells at Charles City, Iowa. Perhaps these features can be shown to have some regional significance.

Prokopovich (1955) includes "brassy" oölites under the generic term corrosion zone, although the formation of oölites is a process of accretion rather than corrosion. Beds of "brassy" oölites and corrosion zones would seem to be related only in the probability that they both formed in environments that were receiving little sediment of any sort.

The formation increases in thickness northwestward across Fillmore County from 22 to 46 feet and continues to thicken northwestward until it is nearly 90 feet thick at St. Paul, Minnesota. It consists of about 30 feet of shale

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PLATE 4.—GALENA AND DUBUQUE EXPOSURES

FIGURE 1.—Stereogram of Galena-Dubuque-Maquoketa strata in Gill's Quarry (F217). Quarry in north bank of South Branch of the Root River (SE¼ SE¼ sec. 17, T. 102 N., R. 12 W.). Inaccessible face is about 90 feet high from the base hidden behind trees. Gradation from massive Stewartville to typical Dubuque rock shows clearly at about the middle. Top few feet (covered by bushes at left) is yellowish basal beds of the Maquoketa Formation. The feldspathized shale of the Dubuque is in the shale under the conspicuous overhang at right.

FIGURE 2.—Cummingsville section (F164). Type section of the Cummingsville Member of the Galena Formation. The boy is standing about 2 feet above the base of the member. (SE¼ sec. 21, T. 105 N., R. 12 W.)

FIGURE 3.—Mystery Cave (F143). The contact of the Dubuque and Maquoketa formations is under the overhang and 2 feet above the yardstick. Exposure at the cave entrance where the river goes underground (SE¼ sec. 19, T. 102 N., R. 12 W.).

and 15 feet of interbedded limestone and shale in the vicinity of Decorah and grades southeastward from there into limestone and ultimately to dolomite.

A few phosphatized snails of the sort characteristic of the basal Guttenburg limestone (Agnew, 1948, Ph.D. thesis, Stanford Univ.; Kay, 1935a, p. 290) have been found loose in the shale near the base of the Decorah at two localities. This suggests that the top of the Spechts Ferry may be locally identifiable in the area where the Guttenburg has graded to shale. However, in Fillmore County, north of there, and even into Glenwood Township of Winnishiek County, Iowa, the Spechts Ferry criteria are not sufficiently conspicuous to justify the use of the name. It is probable that the Spechts Ferry base in Iowa is continuous with the base of the Decorah of this report in Minnesota. The concentration of *Glyptorthis bellarugosa* that marks the base of the typical Ion Member is not present in Minnesota.

The *Prasopora* zonule at the top of the Ion Member in the type area and elsewhere in near-by Iowa and Wisconsin is present over the top of the shale section called Decorah. Because of the predominance of carbonate in the rock of this zonule it is considered in this report to be the basal part of the Galena Formation. The evidence is good, then, that the Decorah Formation of this report is contained between the same horizons as are the three members in their type areas, and that these members are indistinguishable in the region of Decorah, Iowa, and north of there.

Kay (1954) places a disconformity between the Platteville and Decorah formations, but the writer found no physical evidence of it. By inference the Carimona Member is latest Black River in age (Kay, 1954, p. 284), which puts the level of most significant faunal change at the level of the change from limestone to shale; or at the first significant change of environment above the St. Peter-Glenwood beds.

Galena Formation

General.—Hall (*in* Foster and Whitney, 1851, p. 146) proposed the name Galena for gray and drab friable limestone lying above the "Trenton" (Platteville and low Decorah) at numerous localities in Illinois, Iowa, and Wisconsin. He referred (p. 147) to a "... still higher, thin-bedded, argillaceous limestone ..." which apparently he considered to be a distinct member of the Galena and the topmost unit of the Ordovician. Because he believed the Galena limestone to underlie the "Hudson River beds" (modern Maquoketa) it is probable that Hall intended to include what we call Dubuque in the Galena Formation.

The Galena Formation in southeastern Minnesota comprises the carbonate strata that overlie the Decorah shale and underlie the evenly interbedded limestone and shale of the Dubuque Formation. It consists of thin, crinkly beds that are bonded into thick ledges separated by conspicuous bedding planes, a character that gives the formation a thick-bedded gross appearance, especially if fresh. The formation is subdivided into three members: the Cummingsville interbedded limestone and shaly limestone, the Prosser limestone, and the Stewartville dolomitic limestone. The thickness of the Galena is 205 feet in the Spring Valley well and 198 feet at Mahood's Creek. A number of considerations indicate that the formation is nearly 200 feet thick on the average over northeastern Iowa, although it is somewhat thinner farther southeast. North of Fillmore County it is nowhere completely exposed.

Cummingsville Member.—The Cummingsville Member (Weiss, 1955) is an argillaceous limestone, yellowish gray to pale yellowish brown, generally microgranular, thin and crinkly bedded, with conspicuous to thick shale partings. Thin beds are typically grouped into massive units that are separated by smoother, more conspicuous shaly bedding planes. The amount of detrital material varies cyclically throughout the unit, so that, upon weathering, alterante units of shaly limestone and limy shale are developed and give the outcrop a serrated profile (Pl. 4, fig. 2). With the advance of weathering the recessed units become thicker at the expense of the ledges. The shaly limestone and limy shale intervals differ principally in the amount of shale in the partings and in the closeness of the partings. This "sawtooth" character extends from within 8 feet of the base to within 15 feet of the top of the member at Cummingsville, but toward the southeast this character is progressively restricted from the base upwards. There are no chert nodules or corrosion zones at the type locality. Southeastward from there a few zones of chert nodules appear in the lower part of the member and above the middle. Two corrosion zones have been found midway between the cherty zones at one locality in southeastern Fillmore County, but nothing more is known of their distribution.

The Cummingsville is not notably fossiliferous. *Resserella* and *Sowerbyella* are the dominant fossils, but they are not so abundant here as they are in the Prosser Member. *Receptaculites oweni* occurs throughout the member except close to the base. This is the lower *Receptaculites* zone.

The base of the Cummingsville is the level at which the shale of the Decorah Formation is succeeded by rock that is predominantly limestone, usually beds of lumpy or nodular limestone in a shale matrix. A thin coquina of *Prasopora* and other bryozoans occurs locally near the top of the 1- to 3-foot interval of nodular limestone and shale at the base of the member. *Prasopora* also occurs in less abundance elsewhere in these nodular beds. This *Prasopora* concentrate in southeastern Minnesota is undoubtedly related to the *Prasopora* bed that marks the top of the Ion Member of the Decorah Formation in Iowa (Kay, 1929, p. 656). The "brassy" oölites typical of the topmost Decorah toward the northwest are rare in the basal Cummingsville beds there.

At the top of the member the level of greatest change from a high detrital fraction to a low one (Cummingsville to Prosser) is closely approximated by a thin bed of sandy and silty limestone that is poorly fossiliferous, contains some phosphatic debris, and locally is very pyritic. This thin bed is arbitrarily designated the topmost bed of the member.

The Cummingsville argillaceous limestone is readily distinguished from the Decorah shale. It is distinguished from the Prosser Member principally by a detrital content at least twice as great as that of the Prosser. The response to weathering is characteristic and distinct from that of the Prosser, and even on fresh faces the higher detrital content is identified by thicker shaly streaks and the earthy luster of the rock.

The insoluble fraction of the Cummingsville Member ranges from 5 to 25 per cent, depending upon whether the samples are from the limestone or very shaly units; it averages 10–15 per cent. Yellowish-brown clay is most abundant, but tiny cubes and aggregates of pyrite also occur. Conodonts and other phosphatic material are scarce. Siliceous silt and very fine sand are commonly present, but in smaller relative amounts than in the Prosser Member. Silicified fossil debris is not as prevalent as in the Prosser, in fact is generaly absent where the Cummingsville is not cherty. The clastic calcite matrix with abundant fossil debris contains very minor amounts of dolomite.

The member is 63 feet thick at the type locality and essentially the same thickness elsewhere in southeastern Minnesota. The extent of the member is not known, but it occurs throughout southeastern Minnesota and northern Iowa. It is not recognizable in the Platteville mineral district (A. F. Agnew, 1955, personal communication).

Prosser Member.—The first reference to the term Prosser was by Winchell and Ulrich (1897) when they diagrammed and described (p. lxxxv–lxxxvii) a section in the Galena limestone from "Prosser's ravine near Wykoff, Fillmore Country". This section is now known to be in the valley of Mahood's Creek, and the real Prosser's ravine exposes only 25 feet of the uppermost Stewartville rock (Pl. 3). Winchell and Ulrich's section is similar to Part *F* of the Mahood's Creek section (Pl. 3) of this report, except that the lower 37 feet is no longer accessible above the mouth of Mahood's Creek. The member name was given by Ulrich (1911a, p. 257), who clearly intended "Prosser" to be a formational name for a faunal zone. (*See* also Ulrich, 1911b, Fig. 7, p. 369, 524, 525.)

Stauffer and Thiel (1941, p. 86) were the first to attempt to trace the origin of the name Prosser and arrived at the same conclusions as did the writer. Accordingly they described (p. 87) a new measured section from the valley of Mahood's Creek at Wykoff, although they still referred to it as Prosser's ravine. Accumulated errors of location and naming of the locality are pointed out here merely to clear the record but not to change the name of the member.

From about 180 feet of limestone bounded by *Prasopora* below and the upper *Receptaculites* zone above, the Prosser is restricted in this report to the limestone lying between very argillaceous limestone below and dolomitic limestone above. Both of these boundaries are gradational in Fillmore County and vicinity and are delineated arbitrarily by detailed features described below. The type section includes a good exposure of the member as restricted but no longer exposes its base. The concept of the type section is here expanded to include all exposures listed as parts of the Mahood's Creek measured section (*F*192), and these include exposures of the entire member within a distance of half a mile.

Definition of the Cummingsville Member (Weiss, 1955), and restriction of the name

Prosser to the upper part of the Prosser of authors may be an invitation to confusion merely for the sake of not adding a new member name to the Galena Formation. However, toward the southeast, where the lower Galena cannot be recognized as Cummingsville, the middle Galena is also unrecognizable as Prosser. The Cummingsville is therefore not a phase or facies of the Prosser. The name Prosser, whether restricted or not, is applicable only over a relatively small part of the upper Mississippi Valley. To give a new name to the strata in the restricted Prosser of Weiss (1955) and the present report would require suppression of the name Prosser. It seems better that the name Prosser not be suppressed.

The limestone of the Prosser Member is yellowish to light olive gray, generally microgranular, thin and crinkly bedded, with very thin shaly partings. Like other members of the Galena, the thin beds are grouped into massive ledges separated by smooth, conspicuous bedding planes (Weiss and Bell, 1956, Fig. 14). A group of several conglomeratic corrosion zones occurs at the top. A number of sandy and silty fossiliferous streaks occur, particularly in the upper part, but apparently become more numerous lower in the member toward the southeast. An interval containing a group of nodular chert layers occurs above the middle of the member in the west-central part of the county. The chert layers become more numerous and also appear in the lower part of the member toward the southeast. The Prosser is magnesian in the upper part, particularly in the sandy-silty layers and corrosion zones. In this respect it grades into the Stewartville Member above.

The Prosser Member is the most fossiliferous unit of the Galena Formation. It contains many species, and some of them are abundant in individuals. Its fauna is dominated strongly by *Sowerbyella minnesotensis* (Sardeson). Most of the fossils are concentrated in streaks and associated with clay, silt, and some very fine sand. In these fossiliferous streaks are many silicified fossils and particularly fossil fragments. Single specimens of *Receptaculites oweni* have been found in the upper Prosser at three localities, and several in the basal few feet at one locality. *Ischadites iowensis* (Owen) is the characteristic sponge of the Prosser Member and is found in the middle and upper parts.

The basal few feet locally weathers like the Cummingsville. The upper contact is gradational in color and amount of dolomite. The contact is therefore arbitrarily set at the top of an interval of silty and conglomeratic limestone and corrosion zones.

Acid-insoluble materials (olive-gray clay, silicified fossil debris, silt, and pyrite) average 5 per cent of the rock but make up 10–20 per cent of the fossiliferous detrital layers. In these same layers the silicified fossil debris is usually dominant; the quantity of silt and very fine quartz sand is less. Collophane is rare in the Prosser. The clastic calcite matrix is crowded with fossil scraps and shows scattered grains of quartz, feldspar, and tourmaline in thin section. Dolomite rhombs are rare except near the Stewartville, especially in the fossiliferous silty limestone layers and corrosion zones. Chalcedonic chert replaces some fossil material.

Although chert occurs in the Cummingsville Member in Minnesota, it is more typical of the Prosser. The features of the Prosser cherts are in every respect similar to those of the Cummingsville.

Chert is scattered on planes or in beds of smooth, rounded, irregularly shaped, white to light-gray porcelanous nodules that are for the most part unfossiliferous, even where closely associated with coquinoid layers. Most nodules contain tiny brown or gray blebs that may be organic in origin; a very few contain calcareous fossils, and some contain silicified fossils.

The nodules are smooth and closely appressed to the limestone matrix. Where associated with corrosion surfaces or fossiliferous detrital layers, they apparently occupy persistent bedding planes. Otherwise they show no preference for bedding planes, unless, in the case of continuous beds of nodules, the bedding planes have been obliterated. The nodules may be mixed with very fossiliferous compact very fine-grained limestone in which the fossils show no silification, or many fossils scattered through the body of the limestone may be wholly or partly silicified. The nodules show no structures or patterns of color suggestive of preserved primary sedimentary structures. Some show fragmental textures relict from the fragmental limestone of their matrix. There is no evidence that the laminae of the limestone matrix follow the contours of the nodules. The bottoms of the nodules are not flatter than the tops; in fact most are uniformly rounded. Where the nodules are associated with fossiliferous detrital streaks the chert grades through "cotton rock" to the limestone matrix. In thin section the "cotton rock" is seen to be normal limestone matrix with incipient chertification shown by many small patches of chalcedonic chert.

Where chertification is more nearly complete the chert contains islands of calcite and isolated rhombs of dolomite.

The number and thickness of chert beds and the continuity of the nodules within beds all increase southeastward toward the mineral district. Chert nodules also become more numerous in the Cummingsville Member in the same direction. An analysis of the criteria of origin of chert listed by McKee (1938, p. 77–84) that are applicable to Galena cherts led to the confusing conclusion that these cherts were both primary and secondary! The writer is inclined to believe that the accumulation was primary, or penecontemporaneous, and that the replacement features are the result of diagenetic effects. Middle Ordovician volcanism would be a reasonable source for the silica. Agnew (1955, p. 1748) considers part of the chert in the zinc-lead district to be of epigenetic origin, which accords with the increase of chert toward that district from Minnesota.

The Prosser Member is 51 feet thick at the type locality and thins to about 40 feet north and west of there. Control toward the southeast is poor, but the member appears to remain about 50 feet thick in that direction. The distribution of the member is very imperfectly known; much of the upper Galena is either gone or covered in the Rochester area, and this is also true in northeastern Iowa. Furthermore, this part of the column is poorly exposed toward the mineral district, so that the place where the Prosser loses its identity southeastward is also not known. On the basis of chert content and relation to the upper *Receptaculites* zone the Prosser correlates with the upper 32 feet of the cherty unit of Agnew (1948, Ph D. thesis, Stanford Univ.; 1956), and perhaps with the basal part of the noncherty unit. Whether the Prosser corrosion zones are present southeast of Fillmore County is still unknown.

Stewartville Member.—The term Stewartville was first used in the literature by Winchell and Ulrich (1897, p. ci) when, referring to the *Maclurites* bed of Sardeson (1892), they said, "This entire bed is exposed in Prosser's ravine near Wykoff . . . and the lower layers are quarried at Stewartville. . . ." Stauffer and Thiel (1941) traced the history of the name Stewartville and concluded that the Stewartville quarry is the type section by original intent. Their concept of the Stewartville is the same as that of Kay (1935b), but although they give the same order of thickness for the unit in Minnesota (50–60 feet) they record (p. 88) *Maclurites* and *Receptaculites* 20 feet lower in the section at Wykoff than did Kay. Kay's data (1935b) concerning the Stewartville in southern Minnesota agree closely with the writer's observations, except that the type section possibly is higher in the column than Kay and earlier workers believed.

The Stewartville Member in a dolomitic limestone, pale yellowish gray to grayish yellow, which weathers to yellowish orange. On the whole it is very fine-grained, microgranular below, and fine-grained or very fine-grained in the upper part. The bedding is thin and crinkly; very little shale is in the partings. Like other members of the Galena Formation the thin beds are grouped into massive units that are parted by conspicuous smooth bedding planes. Where fresh the rock is mottled by irregular patches and blotches that are slightly darker than the body of the rock and seem to contain a higher fraction of dolomite. This mottling is developed especially in the upper half to two-thirds of the member. The mottled patches leach out more rapidly than the body of the rock and leave weathered surfaces (especially on joint planes) pitted in the pattern of the mottling. The appearance of these pitted surfaces is rather like that of Beales (1953, Fig. 3).

The sparse Stewartville fauna is dominated by snails and cephalopods, and most of these occur as steinkerns that contain crinoid columnals and other organic debris. The upper 20–30 feet is crinoidal, and the beds immediately underlying the Dubuque are especially so, as are the basal Dubuque beds. The upper *Receptaculites* zone is poor in numbers of specimens, and consequently it is difficult to be accurate regarding its distribution. The zone extends from about the middle of the lower half to a level somewhat above the middle of the member. The rare specimens in the upper Prosser Member are not considered a part of this zone. The *Palaeosynapta flaccida* zone (Weiss, 1954a) occurs in the upper part of the member. Its thickness and distribution are imperfectly known, but it appears to be thin and to be contained within the limits of 6 and 35 feet below the top of the member. There is some indication that the zone rises stratigraphically southward into Winneshiek County, Iowa.

A thin bed of yellow feldspathized shale (Weiss, 1954c) occurs from 10 to 20 feet above the base of the Stewartville and approaches closer to the base from northwest to southeast. The lower 10–20 feet of the member, between

the base and the feldspathized shale, is thinner-bedded and more argillaceous than the rest of the member. This lower part weathers to a dusty surface because of the clay content, its fresh color is yellower than the Prosser, and its dolomite content is much greater than that of the Prosser. At the top the Stewartville grades into the Dubuque in color, grain size, bedding, dolomite, and insoluble content (Pl. 4, fig. 1). The upper Stewartville is grayish yellow to yellowish orange, fine-grained, uniformly thin and crinkly bedded, very dolomitic, and almost wholly without shale in the partings. Upward the rock gradually becomes grayer, finer-grained, less dolomitic, and the limestone beds become thicker and smoother and are separated by shale beds. The transition takes place through 10–12 feet of rock. The writer draws the contact at the base of the lowest conspicuous shale parting in this transition zone.

The Stewartville contains less insoluble material than any other of the units: 2–3 per cent through most of the member and only 7 per cent in the lower, argillaceous beds. Most of it is yellowish-brown clay, with a little siliceous silt and silicified fossil debris. Collophane is rare, and pyrite is uncommon. The clastic calcite mosaic shows abundant organic fragments in thin section, as well as clots of subhedral dolomite crystals and scattered euhedral rhombs.

The Stewartville is completely exposed in southeastern Minnesota only at Mahood's Creek and Rifle Hill, near Greenleafton. The member is about 85 feet thick at Mahood's Creek and at least that thick farther northwest but thins southeastward to 75 feet at Rifle Hill (Weiss and Bell, 1956). Possibly it is somewhat thinner in Niagara Cave at Harmony, Minnesota, but the deposits on the walls of the cave largely obscure the rock and make this conclusion uncertain. However, these data do accord with the trend of thinning reported by Kay (1935a, p. 292).

Dubuque Formation

The Dubuque Formation was named by Sardeson (1907, p. 193) and was described as, "the strata of irregular limestone and interlaminated carbonaceous shales, which extend at Dubuque, Iowa, from the 'cap rock' to the blue shales of the Maquoketa proper . . . the formation coincides with the *Triplecia* [*Oxoplecia*] bed. . . ."

Kay (1935b, p. 571) restricted the type Dubuque to the thin ledges of carbonate separated by shale and substituted a rock unit for Sardeson's mixed faunal and lithologic unit. Kay's concept of the type section is preferred. At the type section the interval of transition from Stewartville rock to Dubuque rock is thicker than it commonly is in Minnesota. Nevertheless, the Dubuque of this report conforms closely to that of Kay at the type section. The writer has not found the Depauperate zone above the Dubuque in Fillmore County, and evidently this classic marker of the Maquoketa base is not present at the northern fringe of its outcrop. The writer believes that the lateral uniformity and lithologic distinction of the Dubuque warrant its classification as a formation with rank equal to the Galena and Maquoketa.

In Fillmore County the Dubuque is a unit of interbedded limestone and shale. The limestone beds are light olive gray and grade to light brownish gray downward and yellowish gray upward. They are microgranular to very fine-grained, contain little insoluble material, and are magnesian or dolomitic near the top and base of the member. Although the limestone layers are thin and rippled near the base, they are typically even and of medium thickness throughout most of the unit. (*See* Weiss and Bell, 1956, Fig. 15.) Between individual limestone beds are beds of yellowish-gray to light-olive gray shale that, with few exceptions, are somewhat thinner than adjacent limestone beds.

A conspicuous bed of feldspathized shale (Weiss, 1954c) is present in the lower or middle part of the Dubuque (from 12 to 20 feet above the base) and dips regularly northwestward with respect to the base of the formation. Four localities give stratigraphic control on the bed and pieces of it can be found in the float at three additional places in Fillmore County. There is a second feldspathized shale bed 16 feet above the first, but this bed has been observed only at Mystery Cave, near Spring Valley.

The Dubuque is distinguished from the Stewartville by the presence of shale partings and beds. Dubuque limestone is grayer, finer-grained, and less magnesian than that of the Stewartville. The Dubuque is moderately fossiliferous. The change from the poorly fossiliferous Stewartville is pronounced, although the basal Dubuque is itself not particularly fossiliferous. The fossils in the Dubuque are concentrated in some of the shale layers and in the top of the lower feldspathized shale. Near the base the rock is crinoidal and contains local lenses of "criquina".

The Dubuque is distinguished from the overlying Maquoketa by the distribution of the detrital fraction in the carbonate. The shale of the Dubuque is almost wholly confined to beds between the limestone layers (Pl. 4, fig. 3). In the Maquoketa the clay is uniformly destributed throughout the rock to the extent that bedding planes are scattered and poorly developed. Also, the Maquoketa is dolomitic, particularly in the lower part. The change from Dubuque to Maquoketa rock is gradational through a short interval, and the contact is placed at the top of the highest conspicuous shale parting in the succession (Pl. 4, fig. 3).

Most of the acid-insoluble material of the Dubuque is yellowish clay and makes up about 10 per cent of the rock. Silicified fossil debris and pyrite are locally present. Collophane is a minor constituent but is more abundant than in the Stewartville. The clastic calcite mosaic shows numerous fossil fragments and crinoid columnals in thin section, as well as scattered chips of collophane, small streaks and patches of clay, and bits of fossils replaced by chert. Dolomite, where present, occurs as euhedral rhombs replacing the calcite matrix.

Control on the thickness of the Dubuque in southeastern Minnesota and adjacent Iowa is meager, but it is about 34 feet thick at two outcrops and very close to that in the Spring Valley well. The outcrops in Fillmore County are close to the eroded edge, but the formation extends to Dubuque, Iowa, and the mineral district. Although the carbonate beds change from limestone to dolomite toward the southeast, the Dubuque is the unit most uniform laterally, among those studied, in its principal characteristic (interbedded carbonate and shale).

Stauffer and Thiel (1941, p. 91) thought that the correlation of the Dubuque was uncertain and that it might well fall into either the Middle or Upper Ordovician. They called it Upper Ordovician, because most of the diagnostic Galena fossils are absent from the Dubuque. Although the Galena and Dubuque faunas are taxonomically distinct they are not necessarily so from the evolutionary standpoint. The faunal difference may well be due to the different sedimentary environments under which the formations were deposited. The Stewartville fauna consists almost wholly of snails and cephalopods, the Dubuque fauna of brachiopods. Stauffer and Thiel did not take account of the relationship of the Dubuque and Prosser faunas, which is at least as close, if not closer, than the relationship of the Dubuque and Maquoketa faunas.

The fauna is heavily dominated by brachiopods, most of which are intermediate between strictly Galena types and younger ones. The sowerbyellids change within the formation from the *Sowerbyella minnesotensis* type to one closely resembling the big *S. recedens* of the Maquoketa. Among the strophomenids the difference between *Strophomena trilobata* of the Stewartville and *S. occidentalis* is not great, and the latter is found in both the Stewartville and the Dubuque. *Tetraphalerella planodorsata* is a new and conspicuous Dubuque species but is not far removed morphologically from a single specimen of *Strophomena* cf. *S. magna* found in the Stewartville. Among the rafinesquinids, *Rafinesquina sardesoni* is an intensification of the form seen in *R. camerata*, and small individuals are very much like the latter. *Rafinesquina kingi* is typical of the Maquoketa, and only one specimen was found in the Dubuque. *Megamyonia unicostata* is present in the Dubuque in some numbers and is a new form indicating affinity with the Maquoketa. *Resserella corpulenta* grades upward through the Dubuque from a generally small form like *R. rogata* of the subjacent rocks to the generally large form typical of the upper Dubuque and the Maquoketa. *Oxoplecia ulrichi* is new and distinctive in the Dubuque but suggests no affinities with the rocks either above or below and, incidentally, is so rare that it is worthless as a zone marker.

The foregoing evidence, added to that of earlier authors, shows clearly that the Dubuque fauna is transitional, but the writer believes that there is a balance in favor of Galena affinity. The most characteristic feature of Upper Ordovician brachiopod faunas is the abundance of large orthids with numerous ribs. These are a conspicuous part of the Maquoketa fauna of Minnesota, although they are usually outnumbered there by *Sowerbyella* and *Resserella*. Large, costellate orthids are conspicuous for their rarity in the Dubuque; only one specimen each of *Austinella kankakensis* and *Plaesiomys* cf. *P. proavita* has been found in collections from nearly a dozen different localities. These facts are the basis for considering the Dubuque to be more closely related faunally to the Galena and therefore uppermost Middle Ordovician in age.

Maquoketa Formation

The Ordovician rocks above the Dubuque Formation were not studied critically and are lumped under the name Maquoketa without differentiation into members. The relation of these rocks to the members of the Maquoketa in Iowa is not known, and the relation of Sardeson's Wykoff Formation to the body of the Maquoketa is uncertain. Modern exposures of these rocks in Minnesota are inadequate to demonstrate a two-fold division into Maquoketa and Wykoff formations in the manner of Sardeson (1897), although there is a unit of rock lying on the Dubuque at Granger, Minnesota, that is different from the rocks of the Wykoff-Spring Valley area. Pending critical work on the Upper Ordovician rocks of this area the term Wykoff is held in abeyance.

The Maquoketa in Fillmore County consists of about 70 feet of very argillaceous dolomitic limestone that is massively bedded, generally microgranular, grayish yellow, vuggy in the lower part, breaks with conchoidal fracture, is poorly fossiliferous except in streaks, and weathers rapidly to a shaly slope. At the top is a thin bed with sand and abundant chert and quartz granules. Typical Upper Ordovician fossils were found 14 feet below this conglomeratic bed about a mile west of Mystery Cave (Pl. 4, fig. 3). At the several localities where the conglomeratic bed is exposed, typical rock of the Devonian Cedar Valley Formation (Stauffer and Thiel, 1941, Fig. 34) occurs 10–20 feet above it. The rock immediately above and below the conglomeratic bed is the same. This placement of the Ordovician-Devonian boundary is provisional, and there is at least the possibility that the conglomeratic bed is the base of Sardeson's Bed 14 (1897, p. 33).

Stratigraphic Paleontology

Ranges and Zones

Systematic paleontology done in connection with this study has been reported elsewhere (Weiss, 1954a; 1955). The occurrences and ranges of the species in large, carefully controlled collections[4] are recorded in Table 1. The most fossiliferous rock units yielded sufficient specimens to permit their division into lower, middle, and upper parts for the tabulation of species. The use of *cf.* designations for poorly characterized specimens intermediate between occurrences of well-defined species (*e.g.*, *Pionodema* cf. *P. conradi* in the Carimona Member) emphasizes the gradational continuity of the evolutionary lines.

Distinctive faunal groups commonly have about the same limits as the containing rock units. This correspondence between physical environment and fossil community indicates predominance of environmental control of the occurrences and abundances of the species. This is not to say that the faunas did not change with time, but that evolutionary changes from one bed to the next are not great. There is some indication that some faunal boundaries are oblique to lithologic boundaries, but the disparity is so slight as not to be observed across a single county, or even two.

Several faunal zones are useful supplements to the rock units. The lower *Receptaculites* zone spans all but the lower part of the Cummingsville Member and extends rarely into the basal Prosser. This zone is particularly useful because of the abundance of individuals. The upper *Receptaculites* zone is poorer in specimens and thinner. Its limits are uncertain at some localities, but apparently it occupies the middle and lower middle parts of the Stewartville Member. The rare specimens from the upper part of the Prosser are not included in this zone, for the species seems to be absent from the lower part of the Stewartville.

The range of *Ischadites iowensis* was formerly believed (Weiss, 1953, Ph.D. thesis, Univ. Minnesota) to be largely confined to the middle and upper parts of the Prosser Member but not mixed with the few *Receptaculites oweni* of the basal or topmost Prosser. Contrary to this rule and to Weiss (1954a, p. 428), a single specimen of *I. iowensis* was found with *R. oweni* in the upper Cummingsville at Mahood's Creek, and a cluster of small specimens has been collected from the middle of the Cummingsville, at Cummingsville.

The limits of the zone of *Palaeosynapta flaccida* are as yet imperfectly known (*see* Stewartville Member), but it appears to be an indicator of the upper Stewartville. Hansen (1951, M.Sc. thesis, Univ. Minnesota) made a detailed and precise study of the stratigraphic distribution of ostracods in the Decorah at St. Paul, Minnesota, that suggested that ostracod zones may be distinguished in this thick shale unit.

Black River-Trenton Boundary

Of the index fossils listed by the Ordovician Subcommittee (Kay, 1954) that are identified

[4] Stored in Geological Museum, University of Minnesota.

to species or whose ranges define the Black River-Trenton boundary, almost all are absent or rare in Minnesota rocks. Exceptions are *Doleroides* sp., *Öpikina* sp., and *Flexicalymene senaria*, which are included in the lists diagrammed in Table 2. This table shows clearly that assemblages of typical species and genera of the middle and upper Black River and the lower Trenton lose their identity by mixing and intergrading, if the stratigraphic distribution of these same fossils in Minnesota is considered. Characteristic Black River and Trenton species listed by Swartz (1948, p. 1571–1572) shed no light on the problem, because, except for those already included in Table 2, he lists species not found in Minnesota. It is obvious that the macrofaunal characters of those elusive concepts Black Riveran and Trentonian do not occur in the Ordovician rocks in Minnesota and therefore that the distinction between these concepts is impossible on that basis in Minnesota. Identification of Minnesota rocks as Black River or Trenton is futile, as Sardeson (1892) long ago realized. The time correlation of parts of the Minnesota Ordovician section with the Black River and Trenton of New York on paleontologic grounds appears equally meaningless. If the views of Winder (1953) are correct the reason for the failure of correlation is obvious. Continuous comparison of intervening beds from Minnesota to New York might establish physical correlatives and gross time correlatives, but only these.

Not only is there no sharp faunal break in the Middle Ordovician in Minnesota, but the so-called faunal breaks of New York cannot be made to fit the facts of fossil distribution in the Minnesota rocks. The rather rapid change of numbers, relative dominance, and species of fossils near the Platteville-Decorah boundary has long been considered of time significance. In reality the changes are largely quantitative rather than qualitative and reflect the significant change of bottom environment at that level.

Geologic History

Sedimentary Environments

The formations discussed above were deposited in an embayment of the Middle Ordovician shelf sea lying between the positive areas of the Wisconsin Arch and the Sioux Uplift. In the sea that invaded this region were distributed the St. Peter and Glenwood sand and mud at the base of the Middle Ordovician succession (Dapples, 1955).

The shore line in its initial stages very probably was continuous northwestward from the Wisconsin Arch around the south edge of that part of the Canadian Shield in northern Minnesota. Whether this was continuous with a western shore on the Sioux Uplift is doubtful, for the Ordovician rocks of the Dakotas are similar to those of the upper Mississippi Valley area.

Some workers have assumed that the facies of the Middle Ordovician of the upper Mississippi Valley change along a north-south axis, and that the thickest Decorah section (at Minneapolis-St. Paul) represents the closest approach to the old shore zone. The writer believes that the axis of facies change is really more nearly east-west in Minnesota (Fig. 2), and that the apparent northward increase in the thickness of the Decorah shale is an accident of geographical distribution of outcrops. The empirical axis of lateral change of formation thickness, horizon of bentonite beds, and cherty zones, is a line trending north of west. The presence of the Pecatonica Member in extreme southeastern Minnesota and also in the area of the Twin Cities and Ellsworth, Wisconsin, suggests that the axis of facies change may be a line somewhat south of west. In other words, a thick shale section (grading into limestone) is not *prima facie* the near-shore facies of the deposit but may be, as in this instance, the offshore facies. This view of the paleogeography is in accord with the findings of Berg (1954, p. 876) for the Upper Cambrian of the region. This is not to say that Middle Ordovician paleogeography depends upon that of the Upper Cambrian, but only that in this instance the basin-upland spatial relations appear to have been the same.

The Pecatonica and McGregor members of the Platteville Formation and the Prosser and Stewartville members of the Galena Formation exhibit the characters typical of undathems (Rich, 1951, p. 10). The bedding of these rocks clearly shows agitation of the water and is identical to the type illustrated by Rich (1951, Pl. 1, fig. 2). The rock is a very fine calcarenite, and most of the shale is confined to the partings. In these rocks too the thicker fossils (*e.g.*, corals, brachiopods) are nearly all embedded with their convex sides uppermost.

The shale of the Decorah Formation and the massive argillaceous carbonate rock of the Maquoketa Formation are fondothems (Rich, 1951, p. 13). Both formations are relatively homogeneous, although much of the Decorah is fissile shale. Both are made of clay-size par-

TABLE 1.—STRATIGRAPHIC DISTRIBUTION OF FOSSIL SPECIES IN THE UPPER MIDDLE ORDOVICIAN ROCKS OF FILLMORE COUNTY AND ADJACENT AREAS

The relative abundance of each species within each unit is indicated by symbols.

R—rare
X—present or readily found at most outcrops
A—abundant and/or dominant

Faunal List \ Rock Units	Pecatonica	McGregor	Carmona	Lower Decorah	Middle Decorah	Upper Decorah	Cummingsville	Lower Prosser	Middle Prosser	Upper Prosser	Stewartville	Dubuque	Maquoketa
PORIFERA													
Receptaculites oweni Hall	...	...	...	...	...	...	A	R	...	R	X	...	...
Ischadites iowensis (Owen)	...	...	...	...	...	...	R	...	X	X	...	...	...
Ischadites sp.	...	...	...	..	...	...	...	...	...	...	R	...	...
COELENTERATA													
Conularia sp.	...	...	...	...	...	...	...	...	...	X	X	X	X
Streptelasma corniculum Hall	...	...	...	...	...	...	X	X	X	X	X	...	...
S. cf *S. corniculum* Hall	...	...	...	X	X	...	...	...	...	...	...	...	...
Lambeophyllum profundum (Conrad)	...	X	...	X	X	...	...	...	...	...	...	...	...
L. breve (Winchell and Schuchert)	...	...	...	R	...	...	...	...	...	...	...	...	...
Catenipora rubra Sinclair & Bolton	...	...	...	...	...	...	...	...	...	...	R	...	...
ECHINODERMA													
Pleurocystites sp.	...	...	...	...	...	...	...	...	X	X	...	...	...
Palaeosynapta flaccida Weiss	...	...	...	...	...	...	...	...	...	...	X	...	...
ANNELIDA (?)													
Rauffella filosa Ulrich	...	...	...	X	X	...	...	...	...	...	...	...	...
BRYOZOA													
Proboscina sp.	...	...	...	R	...	...	...	...	...	...	...	R	...
Ceramoporella sp.	...	...	...	X	...	...	...	...	...	...	...	...	...
Favositella sp.	...	...	...	R	...	...	...	...	...	...	...	...	...
Monticuliporella sp.	...	...	...	R	...	...	...	...	...	...	...	...	...
Atactoporella sp.	...	...	...	X	...	...	...	...	...	...	...	...	...
Homotrypa sp.	...	...	X	A	X	X	X	...	X	...	...	...	...
Homotrypella sp.	...	...	...	X	X	...	...	...	...	...	...	...	...
Prasopora sp.	...	...	R	X	X	A	X	X	...	...	...	...	...
Aspidopora sp.	...	...	...	...	...	X	X	...	...	...	...	...	...
Bythopora sp.	...	...	...	R	...	...	...	...	...	...	...	...	...
Eridotrypa sp.	...	...	...	X	X	X	X	...	X	...	...	...	...
Batostoma spp.	...	...	...	A	X	X	X	...	...	...	...	...	...
Monotrypa sp.	...	...	...	...	...	...	X	...	...	...	...	X	...
Hallopora sp.	...	...	...	X	X	...	...	...	...	...	...	...	...
Subretopora sp.	...	...	...	...	X	...	X	...	...	...	...	...	...
Escharopora sp.	...	X	...	X	X	X	...	...	...	...	...	...	...
Stictoporella cribrosa Ulrich	...	...	...	R	...	X	...	...	...	...	...	...	...
S. frondifera Ulrich	...	...	...	A	X	R	...	...	...	...	...	...	...
Stictoporella sp. (ramose)	...	...	...	X	...	...	...	...	...	...	...	...	...
Rhinidictya sp.	...	X	X	X	X	X	X	X	...	...	...	...	...
Phyllodictya sp.	...	...	X	X	X	X	X	X	...	...	...	...	...
Pachydictya sp.	...	X	...	...	X	X	...	...	...	...	...	...	...
Eurydictya sp.	...	...	...	X	...	X	...	...	...	...	...	...	...

TABLE 1.—*Continued*

Faunal List / Rock Units	Pecatonica	McGregor	Carimona	Lower Decorah	Middle Decorah	Upper Decorah	Cummingsville	Lower Prosser	Middle Prosser	Upper Prosser	Stewartville	Dubuque	Maquoketa
BRACHIOPODA													
Lingula spp.	...	X	X	X	...	...	X	X	X	X	...	...	....
L. iowensis Owen	...	...	...	...	...	...	...	...	...	...	X	X	....
L. philomela Billings	...	...	...	...	...	...	...	...	...	...	...	X	....
"Conotreta" sp.	...	...	...	...	...	...	R	...	...	...	...	....	....
Petrocrania sp.	...	...	...	X	X	...	...	...	...	...	...	....	....
Hesperorthis tricenaria (Conrad)	...	X	...	X	X	X	X	X	X	X	...	....	....
Glyptorthis bellarugosa (Conrad)	...	X	...	...	...	...	X	X	X	X	X	X	....
Vellamo americana (Whitfield)	...	...	...	...	...	...	X	X	X	...	...	....	....
Campylorthis deflecta (Conrad)	...	X	...	...	...	...	...	...	...	...	...	....	....
Dinorthis pectinella (Emmons)	...	...	...	...	...	R	...	...	...	...	...	....	....
Dinorthis sweenyi (Winchell and Schuchert)	...	...	...	...	...	...	X	X	X	...	...	....	....
Plaesiomys meedsi (Winchell and Schuchert)	...	...	...	...	...	...	X	R	R	R	...	....	....
P. subquadrata (Hall)	...	...	...	...	...	...	...	...	...	...	...	....	X
P. cf *P. proavita* (Winchell and Schuchert)	...	...	...	...	...	...	...	...	...	...	...	R	....
Austinella kankakensis (McChesney)	...	...	...	...	...	...	...	...	...	...	...	R	X
Plectorthis plicatella (Hall)	...	...	...	...	...	...	R	...	...	R	...	....	....
P. plicatella trentonensis Foerste	...	...	...	...	...	...	X	X	X	X	X	....	....
P. plicatella minnesotensis Weiss	...	X	...	R	...	...	...	...	...	...	...	....	....
Doleroides pervetus (Conrad)	...	...	X	X	X	R	...	...	...	...	...	....	....
D. pervetus gibbosus (Billings)	...	...	...	A	...	...	...	...	...	...	...	....	....
Platystrophia trentonensis McEwan	...	...	...	...	...	...	X	...	...	...	...	....	....
P. trentonensis perplana McEwan	...	...	...	...	...	...	X	...	...	...	...	....	....
P. amoena McEwan	...	...	...	...	...	X	X	X	X	X	...	....	....
P. amoena longicardinalis McEwan	...	...	...	...	...	...	...	X	X	...	...	....	....
P. extensa McEwan	...	...	...	...	...	...	...	...	X	X	...	....	....
Skenidioides anthonense (Sardeson)	...	R	R	...	...	...	...	...	...	...	...	....	....
Camerella panderi Billings	...	...	...	R	...	...	...	...	...	...	...	....	....
Parastrophina hemiplicata (Hall)	...	...	...	...	...	...	X	X	X	X	...	....	....
Oxoplecia ulrichi (Winchell and Schuchert)	...	...	...	...	...	...	...	...	...	...	...	R	....
Rhynchotrema ainsliei (Winchell)	...	...	...	X	X	...	...	...	...	...	...	....	....
Rhynchotrema capax (Conrad)	...	...	...	...	...	...	...	...	...	...	...	....	X
R. increbescens increbescens (Hall)	...	...	...	...	X	X	X	X	X	X	...	....	....
R. increbescens minnesotensis (Sardeson)	...	X	...	X	X	...	...	...	...	...	...	....	....
R. increbescens var. laticosta W. and S.	...	...	...	...	...	...	...	...	R	...	...	....	....
R. aff. *R. increbescens* (Hall)	...	...	...	...	...	...	...	...	...	...	R	....	....
Rhynchotrema erectus (Wang)	...	...	...	...	...	...	...	...	...	...	...	....	R
Protozyga nicolleti (Winchell and Schuchert)	X	X	A	R	...	...	...	...	...	...	...	....	....
Cyclospira bisulcata (Emmons)	...	...	...	...	...	...	...	...	X	...	...	....	....
Zygospira recurvirostris (Hall)	...	...	...	X	X	X	X	...	X	X	...	....	....
Z. uphami Winchell & Schuchert	...	...	...	...	...	...	...	...	...	...	X	....	....

TABLE 1.—*Continued*

Faunal List \ Rock Units	Pecatonica	McGregor	Carimona	Lower Decorah	Middle Decorah	Upper Decorah	Cummingsville	Lower Prosser	Middle Prosser	Upper Prosser	Stewartville	Dubuque	Maquoketa
Sowerbyella punctostriata (Mather)	...	...	X	X	X	X	R	...	...	...	...	...	...
S. minnesotensis (Sardeson)	...	...	...	...	...	...	A	A	A	A	X	X	...
S. recedens (Sardeson)	...	...	...	...	...	...	...	...	...	...	...	...	A
S. cf *S. recedens* (Sardeson)	...	...	...	...	...	...	...	...	...	...	...	X	...
Rafinesquina alternata (Conrad)	...	...	R	X	X	X	X	...	...	...	...	...	...
Rafinesquina camerata (Conrad)	...	...	...	...	...	...	X	X	X	X	X	...	...
R. cf *R. camerata* (Conrad)	...	...	...	...	...	...	...	...	...	...	...	X	...
R. deltoidea (Conrad)	...	...	...	...	...	...	...	...	X	...	...	...	...
R. cf *R. elongata* Salmon	...	...	...	...	...	...	X	...	...	...	...	...	...
R. kingi (Whitfield)	...	...	...	...	...	...	...	...	...	...	...	R	X
R. praecursor Raymond	...	R	...	...	...	...	...	...	...	...	...	...	...
R. cf *R. prestonensis* Salmon	...	...	...	...	...	...	...	...	X	X	...	...	...
R. sardesoni Salmon	...	...	...	...	...	...	...	...	...	...	...	X	...
Öpikina minnesotensis (Winchell)	X	X	...	...	...	...	...	...	...	...	...	...	...
Ö. transitionalis (Okulitch)	...	...	...	X	...	...	...	...	...	...	...	...	...
Öpikina inquassa (Sardeson)	...	...	...	...	X	X	X	...	...	...	...	...	...
Strophomena incurvata (Shepard)	...	...	X	A	X	X	...	...	...	...	...	...	...
S. cf *S. incurvata* (Shepard)	...	R	...	...	...	...	...	R	...	...	...	...	...
S. cf *S. magna* Wilson	...	...	...	...	...	...	...	...	...	...	R	...	...
S. occidentalis Foerste	...	...	...	...	...	...	...	...	...	...	X	X	...
S. trilobata Owen	...	...	...	...	...	...	...	...	...	...	X	...	...
S. cf *S. trilobata* Owen	...	...	...	...	...	...	...	X	X	X	...	...	...
S. trentonensis Winchell and Schuchert	...	...	...	X	X	X	X	...	X	X	...	...	...
Tetraphalerella planodorsata W. and S.	...	...	...	...	...	...	...	...	...	...	R	X	...
Megamyonia unicostata (Meek and Worthen)	...	...	...	...	...	...	...	...	...	...	...	X	X
Holtedaklina emaciata (Winchell and Schuchert)	...	...	...	...	...	R	R	...	...	...	...	...	...
Resserella corpulenta (Sardeson)	...	...	...	...	...	...	...	...	...	...	...	A	A
R. corpulenta ignota (Sardeson)	...	...	...	...	...	...	...	...	...	...	...	...	R
R. cf *R. corpulenta* (Sardeson)	...	...	...	...	...	...	...	...	...	...	R	...	...
R. cf *R. emacerata* (Hall)	...	...	...	...	...	...	...	...	...	...	...	R	X
R. porrecta (Sardeson)	...	...	...	...	...	...	...	...	...	...	...	R	R
R. rogata (Sardeson)	...	...	...	...	X	X	X	X	X	X	X	...	...
Pionodema conradi (Winchell)	...	X	...	...	...	...	...	...	...	...	...	...	...
P. cf *P. conradi* (Winchell)	...	...	X	...	...	...	...	...	...	...	...	...	...
P. subaequata (Conrad)	...	...	X	A	X	R	...	...	...	...	...	...	...
P. cf *P. subaequata* (Conrad)	...	...	X	...	...	...	...	...	...	...	...	...	...
P. circularis (Winchell)	...	...	...	...	R	X	R	...	...	...	...	...	...

TABLE 1.—*Continued*

Faunal List \ Rock Units	Pecatonica	McGregor	Carimona	Lower Decorah	Middle Decorah	Upper Decorah	Cummingsville	Lower Prosser	Middle Prosser	Upper Prosser	Stewartville	Dubuque	Maquoketa
PELECYPODA													
Saffordia sp.												X	
Ctenodonta sp.												X	
Cyriodonta sp.				X				X	X				
C. of *C. parva* Ulrich			X										
Vanuxemia sp.				X									
Ambonychia bellistriata Hall								X	X	X			
Byssonychia sp.									X				
Modiolopsis sp.				X					X			X	
GASTROPODA													
Sinuites sp.				X							X		
Bucania sp.			X		X	X			X				
Tetranota sp.				X									
Salpingostoma sp.							X						
Phragmolites sp.				X	X								
Bellerophon sp.				X									
Loxobucania sp.			X										
Loxoplocus sp.			X	X	X		X		X				
Trochonema (*Trochonema*) sp.							X				X		
Liospira sp.		X		X			X						
Hormotoma gracilis Hall		X	A		X	X	X						
H. bellicincta Hall							X						
H. major (Hall)											X		
Ectomaria sp.							X						
Helicotoma sp.							X						
Maclurites manitobensis (Whiteaves)											X		
M. crassa (Ulrich and Scofield)											X		
M. cuneata (Whitfield)											X		
Holopea sp.									X		X		
Subulites (*Subulites*) sp.										X			
S. (*Fusispira*) sp.							X		X		X		
CEPHALOPODA													
Endoceras sp.			X	X	X								
Metaspyroceras sp.				X									
Actinoceras sp.			X										
Westonoceras minnesotense (Clarke)											X		
TRILOBITA													
Achatella achates (Billings)											R		
Bumastus porrectus Raymond				X		X							
B. billingsi Raymond and Narraway							X						
B. cf *B. billingsi* Raymond and Narraway											R		
Bumastus sp.		X	X						X				

TABLE 1.—*Concluded*

Faunal List \ Rock Units	Pecatonica	McGregor	Carimona	Lower Decorah	Middle Decorah	Upper Decorah	Cummingsville	Lower Prosser	Middle Prosser	Upper Prosser	Stewartville	Dubuque	Maquoketa
Calliops schmidti (Clarke)	...	...	R	...	...	R	...	...	...	...	...	...	...
Ceraurinus sp.	...	...	...	...	...	...	...	...	...	...	R	...	...
Ceraurus pleurexanthemus Green	...	...	...	X	X	X	...	...	...	...	...	...	...
Ceraurus sp.	...	...	...	...	...	...	...	...	...	...	...	R	...
Dimeropyge galenensis (Clarke)	...	...	...	R	...	...	...	...	...	...	...	...	...
Encrinurus sp.	...	X	...	...	...	...	...	...	...	...	...	...	...
Eobronteus lunatus (Billings)	...	...	...	...	...	...	...	...	R	...	...	...	...
Eoharpes sp.	...	...	...	...	...	...	...	...	...	...	R	...	...
Eomonorachus intermedius (Walcott)	...	...	A	A	X	X	X	...	...	...	...	...	...
Flexicalymene senaria (Conrad)	...	...	...	...	...	...	X	...	X	X	...	X	...
Illaenus americanus Billings	...	...	...	...	...	...	X	X	X	...	X	...	...
Isotelus gigas DeKay	...	...	...	...	...	...	X	...	...	...	...	X	...
Isotelus sp.	...	X	A	X	...	...	X	X	X	X	X	...	...
Odontopleura parvula Walcott	...	...	...	...	...	...	...	...	...	...	...	X	...
Sphaerocoryphe robusta Walcott	...	...	...	...	...	...	R	...	...	...	...	...	...
Thaleops cf *T. ovata* Conrad	...	R	...	...	...	...	...	...	...	...	...	...	...
Tretaspis sp.	...	...	...	R	...	...	...	...	...	...	...	...	...
Vogdesia minnesotensis (Foerste)	...	...	...	...	...	...	...	...	X	X	...	...	...
HEMICHORDATA													
Climacograptus sp.	X	X	X	...	...	...	...	X	X	X	X	...	X
Diplograptus peosta (Hall)	...	...	...	...	...	...	...	...	...	...	...	...	X
Diplograptus sp.	...	...	...	...	...	...	...	...	...	...	...	X	...

TABLE 2.—COMPARISON OF DISTRIBUTION OF CERTAIN FOSSILS, NEW YORK TO MINNESOTA

		Typical Upper[a] Rockland Fossils (Lower Trenton) Kay, 1933, Tbl. 1	Typical Upper[b] Watertown Fossils (Upper Black River) Kay, 1933, p. 3	Typical Rockland[c] (Lower Trenton) Fossils Kay, 1937, Pl. 9	Typical Lowville[d] (Middle Black River) Fossils Young, 1943, p. 237	Typical Chaumont[e] (Upper Black River) Fossils Young, 1943, p. 238
DUBUQUE		10, 11		2		
GALENA	Stewartville	8		5, 6		
	Prosser	8, 11			4	3, 4, 5
	Cummingsville	1, 6, 7, 8, 9, 11		5	1, 2, 4	3, 4, 5, 7
DECORAH		1, 2, 3, 4, 5, 6, 7, 8, 9	2, 3	1, 3, 7	1, 2, 4	1, 2, 3, 4, 5, 6, 7
PLATTEVILLE	Carimona	6, 7, 9	1, 2	3	2	6, 7
	McGregor	3, 5, 7, 12		4, 6	2, 3, 5	1, 2, 4, 5, 7
	Pecatonica				3	

a
*1. Batostoma winchelli
2. Bythopora sp.
3. Escharopora sp.
4. Hallopora sp.
5. Pachydictya sp.
6. Prasopora simulatrix
7. Rhinidictya sp.
8. Resserella rogata
9. Rafinesquina alternata
10. Triplecia sp.
11. Flexicalymene senaria
12. Encrinurus sp.

b
1. Actinoceras sp.
2. Endoceras sp.
3. Tetradium sp. (Leray fm.)

c
1. Phragmolites sp.
2. Triplecia sp.
*3. Doleroides sp.
*4. Bathyurus spiniger
5. Receptaculites sp.
6. Maclurites sp.
7. Gonioceras sp.

d
*1. Batostoma winchelli
2. Rhinidictya sp.
3. Öpikina minnesotensis
4. Zygospira recurvirostris
*5. Bathyurus spp.

e
1. Lambeophyllum profundum
2. Rhynchotrema minnesotensis
3. R. increbescens
4. Hesperorthis tricenaria
5. Glyptorthis bellarugosa
*6. Doleroides spp.
7. Pionodema spp.

*Used as criterion both for Black River and Trenton

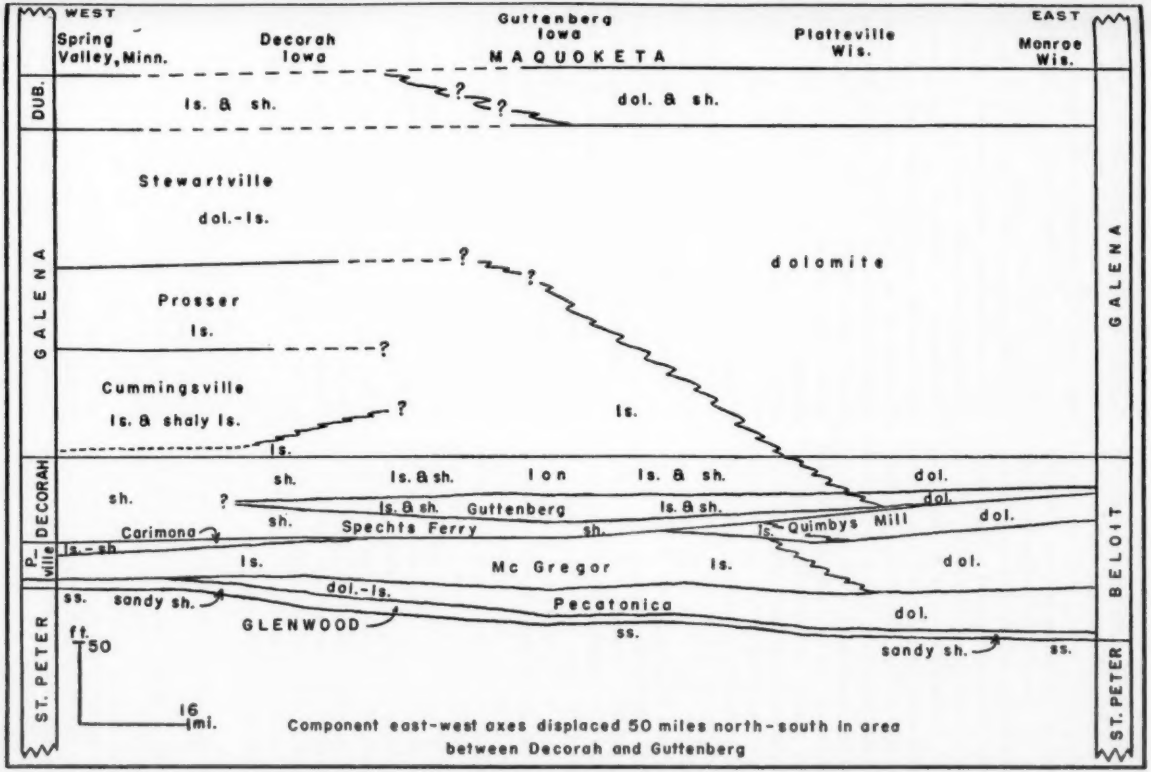

FIGURE 2.—GENERALIZED CROSS SECTION OF FACIES INTERPRETED IN GEOLOGIC HISTORY

Datum is the Decorah-Galena contact. Column at west from this report, column at east and east half of diagram adapted from Figure 34 of Agnew *et al.* (1956). The middle Decorah limy tongue (Guttenberg) pinches out more rapidly farther north. (Myrah quarry, Pl. 5, has about the same longitude as Decorah.)

ticles. Except for the shell beds (see below) the Decorah shale is poor in macrofossils. Most of the fossils of the Maquoketa, however, are scattered ". . .like plums in a pudding. . ." (Rich, 1951, p. 9).

The Carimona Member of the Platteville Formation, the Cummingsville Member of the Galena Formation, and the Dubuque Formaton are tenta tively interpreted as clinothems (Rich, 1951, p. 12) that link the undathems and fondothems. In the major characteristics of the class (alternate bedding of two kinds of rock) they agree with typical clinothems. The interpretation is tentative, however, because search has not yet been made for flow markings on the bedding planes and for evidence of intrastratal flowage. In reality, although these units lie between recognized undathems and fondothems and have some characteristics of clinothems, there is yet no other evidence to suggest that the Ordovician sea bottom did not have a uniform average slope. The Carimona, Cummingsville, and Dubuque were deposited on that part of the bottom slope across which wave base passed as its depth fluctuated. They represent the transitional environment between that persistently above wave base and that normally below it. Unless we can be sure that this transition was coincident with a break in the bottom slope the beds there deposited should not be called clinothems. The terms undathem, clinothem, and fondothem are economical expressions of complex concepts and are useful in this region whether a clinoform actually existed or not. Experience may show that it is helpful in certain instances either to conceive of an undaform grading directly into a fondoform or to broaden the meaning of clinothem to include those clinothemlike strata deposited on a slope that was continuous with the undaform and fondoform.

Sequence of Events

The upper Middle Ordovician rocks of the upper Mississippi Valley appear to record the following sequence of events.

(1) Continued northwestward transgression of the sea followed St. Peter and Glenwood deposition; recorded by the Pecatonica-McGregor undathem. The thickness of the Pecatonica was controlled more by irregularities on the St. Peter-Glenwood surface than by direction of transgression.

(2) Deepening of the sea or downwarp of the shore regions caused the Carimona "clinothem" to advance over a surface of very small average slope and to be followed quickly by the Decorah fondothem. Either the slope of the bottom was so slight that the unda and fondo environments were not widely separated, or the transgression was rapid with respect to thickness of rock accumulated. Whichever may have been the case this "clinothem" is very thin.

(3) Deposition of the Decorah shale in the fondo environment recorded a period of stillstand and succeeding regression of the sea. Because of the gradation to "clino" deposits above and below, and the rather rapid eastward thinning of the Decorah, it appears that the fondo environment invaded southeastern Minnesota rather briefly at this time. The shale mass grades eastward to alternating shale-limestone "clino" beds and then to limestone unda beds. The possible significance of the shell beds in the Decorah shale is considered below.

(4) Bottom environments migrated southwestward in southeastern Minnesota as the sea retreated or shallowed and the Cummingsville "clinothem" developed. The limestone and shaly limestone that are alternately bedded are not as conspicuously distinct as in the Carimona and Dubuque, nor as in typical clinothems (Rich, 1950; 1951). Furthermore the wavy bedding in much of the Cummingsville indicates that the bottom was not free of wave agitation. Most of the member grades eastward within Fillmore County to a limestone undathem. Typical clino characteristics might reasonably be expected only farther southwestward.

The massive ledges of thin wavy-bedded limestone that thicken eastward from the bottom of the member record a brief westward extension of the undaform after the development of the beds at the Decorah-Galena transition.

(5) Regression of the sea, or at least of the sedimentary environments, continued, and the extension of the undaform westward resulted in the deposition of the Prosser and Stewartville members of the Galena Formation.

(6) The change from the regressive condition to renewed transgression apparently is not recorded physically (? corrosion zones at the top of the Prosser?) in the Galena Formation of the upper Mississippi Valley, but the clinothem-like Dubuque Formation implies renewed eastward transgression of bottom environments. The more nearly typical clinothem character of the Dubuque and the distinct fondothem nature of the Maquoketa Formation argue that this marine advance extended much farther over the

positive areas than that recorded by the Decorah Formation. This is in accordance with knowledge of Upper Ordovician seas elsewhere in North America.

(7) Continued transgression of the sea (now Upper Ordovician) brought a typical (Maquoketa) fondo environment to southeastern Minnesota and vicinity. Here the preserved Ordovician rock record ends.

The relations of the several formations of this region are thus seen to be the result of two advances by a shallow shelf sea, separated by a temporary shift of sedimentary environments, perhaps by incomplete withdrawal of that sea. This analysis does not compare with that of Winder (1953) for the Black River, Trenton, and Cincinnatian of New York and adjacent Ontario. This apparent disparity may well be the result of the lack of precise correlations of Minnesota and New York rocks, as discussed earlier.

The writer believes that the three depositional environments were not greatly different in relief and that the angle between time planes (sea bottom surfaces) and at the boundaries of the unda-, "clino-", and fondoforms is very small. These rocks seem to represent a record of marine transgression and regression and resultant deposits that is about as close to "onion-skin" layering of beds as can possibly be distinguished. A corollary of these beliefs is that faunal zones be parallel or very nearly so to the lithologic boundaries. That this is true with minor exceptions has already been shown. The errors warned against by Rich (1951, p. 14) in measuring clinothem sections are insignificant here because of the above, and because of the relative thinness of the "clinothems". This view of a broad shelf sea with three depositional environments not sharply distinguished (and each very broad) is supported by the fact that the bentonite and feldspathized shale beds are in similar host beds over wide areas (but not the whole upper Mississippi Valley) and that these ash layers are only slightly out of parallel with each other (Pl. 5; Weiss and Bell, 1956, Fig. 11).

That the sea was everywhere shallow (*i.e.*, in the zone of abundant light penetration) is evidenced by the profuse bottom faunas of these rocks. This is true even of the Maquoketa fondothem. The high proportion of carbonate rock probably reflects two conditions: (1) diminished supply of coarse terrigenous materials after St. Peter deposition, and (2) effective by-passing of fine detritus to the fondo environment. Distance from shore is not considered to have been an important factor, because the wide undaform should have accumulated coarse detritus to some degree over its surface if such material was supplied at the shore. For example, after formation of the topmost corrosion surface of the Pecatonica a little coarse quartz sand was washed over many square miles of this surface, but no sand occurs in superjacent beds that must have been formed with the shore at about the same position. Thus a little sand, introduced to the basin over a short span of time, was widely spread over the undaform.

Subsidence of the basin probably accounts for the thick deposits, especially those of undaform origin. Whether the regression responsible for the Cummingsville rock was produced by upwarp of the basin or withdrawal of the water is not known, but it was not the result of seaward building of the shore by plentiful detrital deposition. The regression may be illusory, and the undaform and "clinoform" may have grown outward over the Decorah fondothem by mere accumulation of carbonate shell hash, although this appears unlikely on a bottom of very small relief. Certainly the emplacement of the Maquoketa fondothem represents renewed downwarp or rise of sea level or both.

This historical analysis appears somewhat at variance with that proposed by Dapples (1955) and Agnew (1955). Both of them considered the Middle and Upper Ordovician sea of the upper Mississippi Valley region to have moved north or northwestward into Minnesota. The preceding analysis postulates eastward "transgression" in southeastern Minnesota and parts of adjacent states following Pecatonica deposition. This paradox is largely resolved when the scale of the areas studied is considered. Both Agnew and Dapples dealt with large sections of the central states and drew paleogeographic conclusions on a more general basis than does this report. It is not only reasonable, but probable, that a long shore line controlling the conditions that Dapples describes would have irregularities sufficient to produce local variations of the sort described here. As a matter of fact, Dapples' text attributes more significance to the Wisconsin Arch than is indicated by his shore line IV (Dapples, 1955, Fig. 9). The paradox is removed when it is made clear that the events in southeastern Minnesota are not conceived to have been necessarily the result of a shore line sweeping back and forth across southern Wisconsin, but of transgression and regression of bottom environments. Whether exposed or submerged

the Wisconsin Arch would have influenced sedimentation around its borders. The shifts of environments of sedimentation might have been produced by changes of depth of water, warping of the basin or its margins, changes in the terrigenous detritus supplied at the shore, or some combination of these factors. All three must have been effective at different times and in different senses and combinations as the Ordovician sea overspread North America.

Pecatonica and McGregor Thickness

The Pecatonica Member thins to disappearance at its western edge in southeastern Minnesota but is present again farther north. The McGregor Member is locally irregular in thickness in southeastern Minnesota (although it tends to thicken toward the west), especially where the Pecatonica is absent. The northwestward thinning of the Pecatonica appears to suggest northwestward transgression of the carbonate environment but might also be the result of downwarp of the eastern part of the embayment. The writer believes that the cause was the accumulation of thicker parts of the undaform limestone in low places on the irregular depositional surface of the St. Peter-Glenwood.

Corrosion Zones

Weiss (1954b) discussed the diastemic corrosion zones of these rocks, but additional considerations are pertinent to the geologic history of the area. The post-Wisconsin corrosion surfaces illustrated by Hudson (1910) are on bedding planes and are the result of fresh-water etching. Wentworth (1939) studied such benches formed in reef limestone along salt-water shores. He found that the part of the bedrock shore that most nearly resembles corrosion zones in the aimless pattern of ragged pits is the higher, landward side of the shore. This is the part sculptured by fresh-water solution (Wentworth, 1939, Fig. 5; *see* also 1944, p. 123). The greater relief of the pitted surface studied by Wentworth probably results from less resistant limestone, as compared to that of the Ordovician corrosion surfaces. On the seaward side of such pitted areas the salt water retards solution, with the result that there is a limit, set by the level of salt water, below which solution is ineffective. The outer part of the bench is also somewhat smoothed by wave action.

The solution-benching process described by Wentworth cannot account for widespread surfaces like those of the Ordovician. Subaerial or fresh-water-lake origins of these surfaces are also excluded by the lack of karst features and the fossil proof of marine origin of the rock. Submarine corrosion of these surfaces seems assured The anomaly of submarine solution on the bottom of a sea depositing (or having recently deposited) carbonate rocks was not satisfactorily accounted for by Weiss (1954b). Mosebach (1952) has recently shown that waters charged with H_2S can effectively dissolve limestone or lime muds. Corrosion zones did not form by diagenetic solution within the sediment, where Mosebach thinks fossils are dissolved, else they would be stylolitic in nature. They must have formed on the bottom when the water locally and/or temporarily stagnated and H_2S (or CO_2, Prokopovich, 1955) occurred in greater concentration. Pyrite is of course widespread in small concentration in these Ordovician rocks but is especially concentrated at the corrosion surfaces (Weiss, 1954b). The means by which part of an undaform becomes temporarily or locally a reducing environment is unknown but is probably a matter of lack of sufficient agitation of the water. More needs to be known of the extent of individual corrosion zones before this explanation is accepted. The random topography of the corroded surface in a typical corrosion zone (Weiss, 1954b), however, surely precludes Prokopovich's (1955) explanation of origin partly by abrasion by turbid bottom currents.

The use of corrosion zones to mark rock-unit boundaries may appear somewhat inconsistent, inasmuch as they are diastems or small disconformities and very probably also time planes (because of the very short hiatus). They are, however, convenient horizon markers in southeastern Minnesota. The rock units of this report can be recognized, where present, without the corrosion zones. Widespread corrosion zones would of course be advantageous key horizons for relating these rock units to lateral equivalents of somewhat different lithology.

Decorah Shell Beds

The shell beds or coquinoid limy layers of the Decorah shale are of two types. Near the base, particularly in the southeast, there are several layers of soft, punky, light-greenish-gray limy marlstone, usually containing very few fossils. The remaining and more abundant limy layers are hard coquinoid ledges. Some of these contain considerable green shale in pockets or streaks, but most are nearly free

from it. Most of the fossils in the formation are found in these ledges, although there is a scattering of macrofossils throughout much of the shale. Some massive parts of the shale appear to have been deposited so rapidly as to discourage bottom life completely.

Because the shell beds and ledges are so thin with respect to the beds of shale, and because some of them contain shale, it is unlikely that the shell beds represent periods of near or total cessation of clay deposition. Those limy layers near the Carimona and Cummingsville members may have formed near wave base, but regardless of the part of the bottom on which they formed they are clearly in the nature of "lag concentrates". During exceptional storms the water was roiled by fine detritus stirred from the bottom and thrown into suspension. A mat of both newly destroyed and reworked shells and other skeletal material would be left on the bottom. The mat of shells and fragments would be free of clay while the water was still strongly agitated. As the storm passed, the clay in suspension would begin to settle onto and sift through the coquinoid mat of debris. If most of the mat were made of fresh whole shells and other large pieces the interstices would become filled with clay. If, however, there was abundant crumpled debris, perhaps brought from the undaform, the mat would have been tighter and penetrated by less clay. Among Decorah coquinoid ledges those with much clay are made for the most part of whole brachiopod valves and chunks of bryozoans, and those with very little clay are made almost wholly of comminuted organic trash. This same explanation for coquinites of this type has been proposed by Bucher for those of the Cincinnati region (Rich, 1951, p. 5).

There remains only to explain how this process can occur in a presumed clinoform environment, as has been suggested above for the basal and topmost parts of the Decorah Formation. Rich (1951, p. 5) suggests a direct mixing of unda and fondo environments resulting from the lowering of wave base concomitant to severe storms. This may well be true of the Decorah; on a bottom of very small slope the wave base need not have been lowered greatly. The apparent absence of a typical clinoform was discussed above.

Localities and Measured Section

Localities

The localities shown on Plate 1 and diagrammed in Plate 5 are specified below. Unless otherwise stated, each is in Fillmore County, Minnesota. The location of the Mahood's Creek section is described in Plate 3, and its detailed description is in the following section.

Big Spring (F177)—Stream cuts in the N½ SW¼ sec. 3, T.103 N., R. 11 W.

Carter's Creek (F93)—Exposures along Carter's Creek for a distance of about 1 mile in the E½ sec. 9, T. 103 N., R. 12 W.

Cummingsville (F164)—Road cut and quarries from 0.2 to 0.7 miles north of Cummingsville, Olmsted County, Minn., and on the east edge of the SW¼ sec. 21, T. 105 N., R. 12 W. (*See* also Weiss, 1955, p. 765–766; Weiss and Bell, 1956, p. 104–106.)

Highway 16 (F144)—Road cut on U. S. 16, 1.3 miles west of the junction with U. S. 52 north of Preston, and in the middle of the N½ sec. 35, T. 103 N., R. 11 W. (*See* also Weiss, 1955, p. 762.)

North of Fountain (F96)—Road cut on U. S. 52 approximately 3 miles north of Fountain, and in the NE¼ sec. 27, T. 104 N., R. 11 W. (*See* also Weiss, 1955, p. 761.)

South of Fountain (F95)—Road cut and gulch along U. S. 52, 1.5 miles south of Fountain, and in sec. 14, T. 103 N., R. 11 W. (*See* also Weiss, 1955, p. 760.)

Lime City Quarry (F111)—Abandoned quarry and bluffs at the crossing of Deer Creek 2.5 miles north of Spring Valley, and in the NE¼ SE¼ sec. 9, T. 103 N., R. 13 W.

Mabel Quarry (F160)—Quarry and road cut on Minn. 43, 1.5 miles north of Mabel, and at the SE corner of sec. 15, T. 101 N., R. 8 W. (*See* also Weiss and Bell, 1956, p. 97.)

Masonic Park (F201)—Stream, quarry, and road-cut section at the Masonic picnic ground on Minn. 74 at Deer Creek; on the line between SE¼ sec. 11 and SW¼ sec. 12, T. 103 N., R. 13 W.

Myrah Quarry (F86)—Quarry and road cut 1.0 miles southeast of the center of Spring Grove, Houston County, Minn., and in the SE¼ SW¼ sec. 12, T. 101 N., R. 7 W.

Mystery Cave (F143)—Exposures in and around the commercial cave in the east bank of the South Branch of the Root River, in the SE¼ sec. 19, T. 102 N., R. 12 W. (*See* also Weiss and Bell, 1956, p. 100–102.)

Rifle Hill (F171)—Quarries and road cuts in Rifle Hill, west of Canfield Creek, and in the NW¼ sec. 35, T. 102 N., R. 12 W. (*See* also Weiss and Bell, 1956, p. 99.)

Spring Grove Underpass (F85)—Road cut and quarries at Minn. 44 and the railroad, 3.2 miles west of Spring Grove, Houston County, and in the SE¼ SE¼ sec. 17, T. 101 N., R. 7 W.

Sugar Creek (F147)—Quarries and road cut 1.5

miles west of Fountain, on the north edge of the NW¼ sec. 9, T. 103 N., R. 11 W. (*See* also Weiss and Bell, 1956, p. 102–104.)

Waterworks Hill (F149)—Abandoned quarry on Minn. 74 at the east edge of Chatfield; N½ NW¼ sec. 5, T. 104 N., R. 11 W.

Wubbels' Ravine (F168)—Road cut in the ravine in the N½ sec. 2, T. 101 N., R. 12 W.

Measured Section

Space permits only one measured section to be included—the Mahood's Creek section. Because it includes the type Prosser and the Prosser as restricted by Weiss (1955) this newly detailed description is of especial significance. The collections are related to the measured section by combining the locality number and the footage of the base of the fossiliferous bed above the base of the section (*e.g.*, collection F192.159:6 contains the fossils from the bed whose base is 159 feet 6 inches above the base of section F192). Numbers in this system are to the nearest foot, unless the bed concerned is less than a foot thick. Where significant differences in the abundance of specimens in a collection exist, the abundance relative to that of other species in the same collection is expressed as follows: va—very abundant; a—abundant; c—common; no symbol—present but not conspicuous; r—rare.

The Mahood's Creek section (F192) is compiled from a number of exposures in the valleys of Spring Valley and Mahood's Creeks from about the center of the south edge of SE¼ sec. 8, southward to the head of the Mahood's Ravine along the south edge of the SE¼ sec. 20, T. 103 N., R. 12 W., Fillmore County, Minnesota. The location of the section and its parts is shown in Plate 3.

Only rock of the Stewartville Member crop out in Prosser's Ravine, and the type section of the Prosser Member is undoubtedly the series of exposures in the valley of Mahood's Creek. This part of the section has been known for many years and is recorded in several earlier reports. Fossil and rock collections are numbered with respect to the base of the composite section, and the number for the particular subsection is given in parentheses.

	Thickness of unit Feet	Inches	*Footage above base* Feet	Inches
Drift and soil				
Dubuque Formation: 7± feet				
Note: Beds 65 through 68 are recorded from (F192-H).				
68. Limestone; light olive gray, very fine-grained, medium-bedded, with thin to medium shale interbeds, shales squeezed by slumping ledges.	6		248	
67. Limestone; light olive gray, very fine-grained, single thick bed, parted by shales. The exposure at 192-H does not clearly show the Dubuque base, but enough detail can be seen for good correlation with the section west of Prosser's Corner (F165), where the contact is clear. The Galena-Dubuque contact is recorded here on this basis.		9	247	3
Galena Formation: 199 feet				
Stewartville Member: 85 feet				
66. Limestone; yellowish gray, very fine-grained, thin to medium crinkly beds with thin shale partings, very crinoidal, not distinguishably dolomitic in the hand specimen. A slab of crinoidal dolomitic limestone float from the top of Section F (254± feet above section base) contains abundant *Ischadites* sp. not *I. iowensis*.	3	6	243	9
65. Dolomitic limestone; yellowish gray, very fine-grained, rather thin-bedded with ripply shale partings, very crinoidal, fragmental, somewhat burrowed, weathers to smoother surface than beds below.	2		241	9
Note: Beds 48 through 64 are recorded from Mahood's Cave (F192-F).				
64. Dolomitic limestone; pale orange, fine-grained, sugary, heavy ledges of thin crinkly beds with very little shale in the partings, thoroughly burrowed (resulting in pitted weathered surfaces).	32	4	209	5
63. Dolomitic limestone; pale orange, medium-grained, even bed, hard, fragmental, burrowed. *Resserella* sp.		2	209	3

	Thickness of unit Feet	Inches	Footage above base Feet	Inches
62. Dolomitic limestone; pale orange, nearly yellowish gray in the lower 12 feet, generally very fine-grained, heavy ledges of thin crinkly beds with very little shale in the partings, somewhat burrowed, weathers to yellowish orange along joints. *Maclurites* sp. and several *Receptaculites oweni* 3–4 feet below the top and *Hormotoma* sp. 7 feet above the base.	27		182	3
61. Shale; grayish yellow, calcareous, very thinly laminated.		2	182	1
60. Feldspathized shale; pale yellow, thinly laminated, brittle, contains small pods of very pale-orange microgranular to very fine-grained coquinoid limestone. With few fossils in the shale, but most in the limy pods.		3	181	10
Plectorthis plicatella trentonensis,				
Rafinesquina camerata,				
Resserella rogata,				
Sowerbyella minnesotensis (a),				
Strophomena sp.,				
Climacograptus (?), and				
fucoids				
59. Dolomitic limestone; yellowish gray, generally microgranular, thin and crinkly bedded with shaly partings, more argillaceous than the rocks above, weathers to a dusty surface, fragmental, somewhat burrowed. Beginning with this bed the section turns eastward into a gulch leading out of Mahood's Creek.	19	8	162	2
Sowerbyella minnesotensis,				
Hormotoma major, and				
Climacograptus sp. close to the top				

Prosser Member: 51 feet 2 inches

58. Silty dolomitic limestone; yellowish gray, microgranular, thinly laminated, soft, dusty, argillaceous, silty, with thin soft clay streak at base.	8		161	6
Resserella rogata				
57. Conglomeratic dolomitic limestone; medium gray to brownish gray, fine-grained, argillaceous, with some silt.	½		161	5
56. Conglomeratic magnesian limestone; medium dark gray, medium-grained, single irregular, pitted bed, with pebbles of rock like the matrix, but finer-grained.	7		160	10
Resserella rogata and				
Leperditia (?)				
55. Sandy limestone; mottled pale olive and medium gray, fine-grained, sugary, magnesian, hard rough bed, with abundant silty sand. Beds 51 to 58 correspond to the group of conglomeratic silty dolomitic limestones with corrosion zones so well displayed at Masonic Park (F201), but in this scoured gully bottom (F192F) the corrosion zones do not show.	3		160	7
Parastrophina hemiplicata,				
Resserella rogata, and				
scolecodonts				
54. "Fucoidal" limestone; grayish yellow, very fine-grained, thinly laminated, burrowed, argillaceous.	6		160	1
Resserella rogata,				
Sowerbyella minnesotensis, and				
Climacograptus sp.				
53. Conglomeratic limestone; medium gray with patches of grayish				

	Thickness of unit Feet	Thickness of unit Inches	Footage above base Feet	Footage above base Inches
orange, very fine- and fine-grained, somewhat argillaceous, conglomeratic.	1		160	
Resserella rogata				
52. "Fucoidal" and conglomeratic limestone; grayish yellow, microgranular to very fine-grained, burrowed, with conglomeratic center of light-olive-gray fine-grained limestone. The bed is like Bed 54, but the conglomeratic center is like Bed 53. This bed is the highest at F192-B.		6-7	159	5
F192.159:5 (F.25:3) contains				
Parastrophina hemiplicata,				
Zygospira recurvirostris,				
Pleurocystites sp., and				
Climacograptus (?) sp.				
F192.159:6(B.109:3) contains				
Parastrophina hemiplicata, and				
Resserella rogata.				
F192.159 (D.75) contains				
Sowerbyella minnesotensis.				
51. Conglomeratic limestone; pale-yellowish-brown microgranular rock mixed with medium-dark-gray very coarse-grained rock, a lumpy conglomeratic bed, argillaceous, brinoidal, with incipient chertification of some fossils.		2	159	3
F192.159:3 (F.25:1) contains				
Sowerbyella minnesotensis and				
Strophomena sp.				
F192.159:2 (B.109) contains				
Plaesiomys sp.,				
Rafinesquina (?) and				
Resserella rogata and is 4 inches thick.				
50. Cherty limestone; generally yellowish gray or pale yellowish brown, brownish cast of some collections probably due to more advanced weathering; generally microgranular, massive ledges of very thin crinkly beds, with very thin shaly partings, most fossils and crinoid columnals concentrated in discontinuous streaks. With layers of rather closely spaced chert nodules at 4, 5, 6.5, 10.5, and 11 feet above the base of the unit, and with levels of scattered smaller chert nodules at 12, 12.5, and 15 feet above the base. Numerous collections are identified by numbers for the whole section and also for the sublocality.	21		138	3
F192.158 (C.53) contains				
Sowerbyella minnesotensis (a).				
F192.157 (B.107) contains				
Parastrophina hemiplicata,				
Rafinesquina (?),				
Streptelasma corniculum,				
Climacograptus sp., and				
Ischadites iowensis.				
F192.154 (F.20) contains				
Ambonychia bellistriata.				
F192.151 (C.46) contains				
Plaesiomys meedsi,				
Plectorthis plicatella trentonensis,				
Rafinesquina camerata,				
Resserella rogata,				

	Thickness of unit Feet	Inches	Footage above base Feet	Inches
Sowerbyella minnesotensis (a), and				
Pleurocystites sp.				
F192.150 (F.16) contains				
Rafinesquina camerata,				
Rhynchotrema cf. *R. increbescens*,				
Sowerbyella minnesotensis (c),				
Strophomena trentonensis,				
Loxoplocus sp., and				
Subulites (*Subulites*) sp.				
F192.149 (D.65) contains				
Plectorthis plicatella trentonensis,				
Rafinesquina camerata,				
Resserella rogata,				
Rhynchotrema cf. *R. increbescens*,				
Sowerbyella minnesotensis (c),				
Strophomena sp.,				
Vogdesia minnesotensis, and				
Streptelasma corniculum.				
F192.148 (C.43) contains				
Rafinesquina sp.,				
Resserella rogata,				
Sowerbyella minnesotensis (a), and				
Strophomena sp.				
F192.147 (C.42) contains				
Glyptorthis bellarugosa,				
Parastrophina hemiplicata,				
Platystrophia extensa,				
Plectorthis plicatella trentonensis,				
Rafinesquina camerata, *R.* cf. *R. prestonensis*,				
Resserella rogata,				
Sowerbyella minnesotensis,				
Strophomena cf. *S. trilobata*,				
Subulites (*Subulites*) sp., and				
Climacograptus sp.				
F192.146 (F.12) contains				
Rafinesquina camerata.				
F192.145 (C.40) contains				
Glyptorthis bellarugosa,				
Hesperorthis tricenaria,				
Parastrophina hemiplicata,				
Rafinesquina camerata,				
Resserella rogata,				
Sowerbyella minnesotensis (c),				
Eobronteus lunatus,				
Isotelus sp., and				
Sinuites sp.				
F192.144 (E.37) contains				
Parastrophina hemiplicata,				
Rafinesquina cf. *R. prestonensis*,				
Resserella rogata,				
Sowerbyella minnesotensis (a),				
Vogdesia minnesotensis, and				
Pleurocystites sp.				
F192.142 (C.37) contains				

	Thickness of unit Feet	Inches	Footage above base Feet	Inches
Rafinesquina camerata (a), *Vogdesia minnesotensis*, *Streptelasma corniculum*, and *Climacograptus* sp.				
49. Shaly limestone; yellowish-gray shale and light-brown limestone coquina, weathering to a deep notch in the cliff. *Resserella rogata*, *Sowerbyella minnesotensis* (a), and *Vogdesia minnesotensis*		1	138	2
48. Limestone; like Bed 47. This bed is the base of the Mahood's Cave exposure and is limited to 4 feet for that reason. There is no discontinuity between Beds 47 and 48 at Keenan's Cave.	4		134	2
Note: Beds 19 through 47 are recorded from Keenan's Cave (F192-B).				
47. Limestone; yellowish gray to pale yellowish brown, generally microgranular, massive ledges of thin crinkly beds with very thin shale partings, with some patchy streaks of fine- to medium-grained more brownish fragmental coquinoid limestone. *Dinorthis sweenyi* (c), *Glyptorthis bellarugosa*, *Hesperorthis tricenaria*, *Platystrophia amoena*, *Plectorthis plicatella trentonensis* (c), *Rafinesquina camerata*, *Resserella rogata*, *Rhynchotrema increbescens*, *Sowerbyella minnesotensis* (a), *Strophomena* cf. *S. trilobata*, *Vellamo americana*, *Illaenus americanus* from F192E.18 and *Receptaculites oweni* from F192D.33	18	2	116	
46. Limestone; like Bed 23 but less argillaceous than the beds below.	5		111	
Cummingsville Member: 62 feet, 9 inches.				
45. Sandy limestone; medium-dark-gray very fine- to fine-grained hard even bed with columnals and castings or burrow fillings on the top. Contains considerable clay and silty sand. *Sowerbyella minnesotensis*		3	110	9
44. Limy shale; like Bed 35 but with *Receptaculites oweni* (F192D.26).	1	2	109	7
43. Limestone; pale yellowish brown, very fine-grained, 2 even beds parted by a thin shale, harder than the beds below.	6	109	1	
42. Limestone; like Bed 23.		8	108	5
41. Limy shale; like Bed 35, but olive gray.		11	107	6
40. Limestone; like Bed 23.		11	106	7
39. Limestone; medium gray, weathering light brownish gray, fine-grained, even hard bed with wormy top. (F192D.22) *Resserella rogata*, and *Sowerbyella minnesotensis*		2	106	5
38. Limestone; like Bed 36.	3		103	5
37. Shaly limestone; light olive gray, very fine-grained to microgranular, with some fine-grained fossiliferous streaks, less shaly than most such units, thin and crinkly bedded with shale partings. (F192D.17) *Platystrophia* sp.,	3		100	5

	Thickness of unit Feet	Thickness of unit Inches	Footage above base Feet	Footage above base Inches
Isotelus gigas, and *Receptaculites oweni*				
36. Limestone; like Bed 21, but light olive gray.	2	6	97	11
35. Limy shale; pale olive, very thinly bedded, with limy streaks. *Dinorthis sweenyi*, *Platystrophia amoena*, *Rafinesquina* sp., *Resserella rogata* (c), *Rhynchotrema increbescens*, *Sowerbyella minnesotensis*, *Aspidopora* sp., *Phyllodictya* sp., and *Hormotoma gracilis*	1	5	96	6
34. Limestone; like Bed 23. *Receptaculites oweni*	1		95	6
33. Shaly limestone; like Bed 24. *Sowerbyella minnesotensis*, and *Receptaculites oweni*		10	94	8
32. Limestone; like Bed 23 but very fine-grained and very thin-bedded in the top half. Not distinct from Bed 31 at all exposures. With *Receptaculites* throughout. (F192D.10) *Resserella rogata* (a), *Sowerbyella minnesotensis* (a), *Strophomena* sp., *Sphaerocorphe robusta*, and *Pholidops* (?)	5	7	89	1
31. Shaly limestone; like Bed 20. *Receptaculites oweni*	1		88	1
30. Limestone; like Bed 23 with shaly interval 1 foot below top. (F192D.1) *Platystrophia amoena*, *Rafinesquina camerata*, *Rhynchotrema* sp., *Sowerbyella minnesotensis*, *Ischadites iowensis*, and *Receptaculites oweni*	4		84	1
29. Limestone; yellowish gray, very fine- to fine-grained, even bed, hard, burrowed.		4	83	9
28. Limestone; like Bed 23 with very shaly top. With a few streaks of *Sowerbyella* coquina and with *Receptaculites oweni* throughout.	1	6	82	3
27. Limestone; like Bed 21 but more brownish. Has few fossiliferous streaks of very fine-grained limestone with *Sowerbyella minnesotensis*.	4	10	77	5
26. Shaly limestone; like Bed 24.		10	76	7
25. Limestone; like Bed 23.	1	6	75	1
24. Shaly limestone; grayish yellow, microgranular, thin and crinkly bedded, with thick greenish-gray shale partings.		10	74	3
23. Limestone; like Bed 21 but less brown.	1		73	3
22. Shaly limestone; like Bed 20.	1		72	3
21. Limestone; yellowish gray to pale yellowish brown, microgranular, thin and crinkly bedded with thin shale partings, thin beds bonded into a heavy ledge.	1		71	3
20. Shaly limestone; yellowish gray, microgranular, thin and crinkly bedded with thick shale partings, very argillaceous.	1		70	3

	Thickness of unit Feet	Inches	Footage above base Feet	Inches
19. Limestone; very pale yellowish brown, microgranular, massive unit of thin crinkly beds with thin shale partings, somewhat burrowed, weathering to a jagged and hackly gray face. With *Sowerbyella minnesotensis*. This unit extends to the base of the Keenan's Cave exposure, and the cave itself is in the lower part of this unit. The base of this bed is 2 feet above the base of the Galena, the basal beds being exposed on the road above Bed 16.	20		50	3
18. Limestone; like Bed 19.		6	49	9
17. Shaly limestone; yellowish-brown nodular weathered limestone in matrix of shale, with some small rusty specks (probably rotted brassy oölites).	1	6	48	3
Decorah Formation: 42 feet				
16. Covered interval; measured by short level sights up the Wykoff road. Chert beds exposed by the road are correlated with those above Keenan's Cave. At the top of the covered interval grayish-green shale with rusty specks of rotted brassy oölites can be dug out of the ditch.	34	7	13	8
Note: Beds 1 through 15 are recorded from Quinn's Quarry (F192-A).				
15. Shale; like Bed 11.	2		11	8
14. Shell bed; dark-gray very fine- to fine-grained limestone coquina with patches of light-grayish green marlstone. *Doleroides pervetus* (c), *Pionodema subaequata* (a), *Strophomena incurvata*, *Batostoma* sp. *Rhinidictya* sp., *Stictoporella* sp., and *Eomonorachus* sp.		8	11	
13. Shale; like Bed 11.		2–3	10	10
12. Bentonite; yellowish, greasy streak in the green shale.		½	10	9
11. Shale; grayish green, fissile, calcareous.	3		7	9
10. Limy marlstone; light greenish gray, soft, punky.		3	7	6
9. Shale; grayish green, fissile to massive and blocky, calcareous.	1	3	6	3
Platteville Formation: 6 feet, 3 inches.				
Carimona Member: 6 feet, 3 inches.				
8. Limestone; light brownish gray, very fine grained, single even bed, very pyritic, fragmental, with shaly top layer containing *Pionodema* sp.		9	5	6
7. Shale; grayish green, fissile, calcareous, weathered yellowish brown, appears bluish gray when bleached and dry.		6	5	
6. Limestone; light olive to light medium gray, very fine-grained, in 2 even beds with very thin parting, fragmental.		8	4	4
5. Shale; yellowish brown, very thinly laminated, brittle, calcareous, petroliferous, appears bluish gray where weathered and dry. Contains thin layer of coquinoid dark-gray very fine-grained limestone interlaminated with "fucoidal" or wormy argillaceous grayish-yellow-green microgranular limestone at top. The limestone contains *Pionodema subaequata* and *Strophomena incurvata*. The shale contains *Pionodema subaequata* and *Eomonorachus* sp.		7	3	9

	Thickness of unit Feet	Inches	Footage above base Feet	Inches
4. Limestone; yellowish brown, shaly coquinoid, plastered to Bed 3. *Pionodema subaequata*, *Protozyga nicolleti*, *Strophomena* sp., and *Eomonorachus* sp.		1	3	8
3. Limestone; like Bed 1, with thin shale partings and with 2 zones of gastropod molds dividing the unit symmetrically.	2	3	1	5
2. Bentonite; yellowish orange, with thin yellowish-brown calcareous shale at the base.		5	1	
1. Limestone; pale yellowish brown, very fine-grained, fragmental, evenly bedded, extends to the floor of the quarry. The floor is covered with debris so that the rock below this unit cannot be seen, but this unit may be the basal limestone of the Carimona Member.	1			
Floor of the quarry 4 feet above Spring Valley Creek.	254 ft.			

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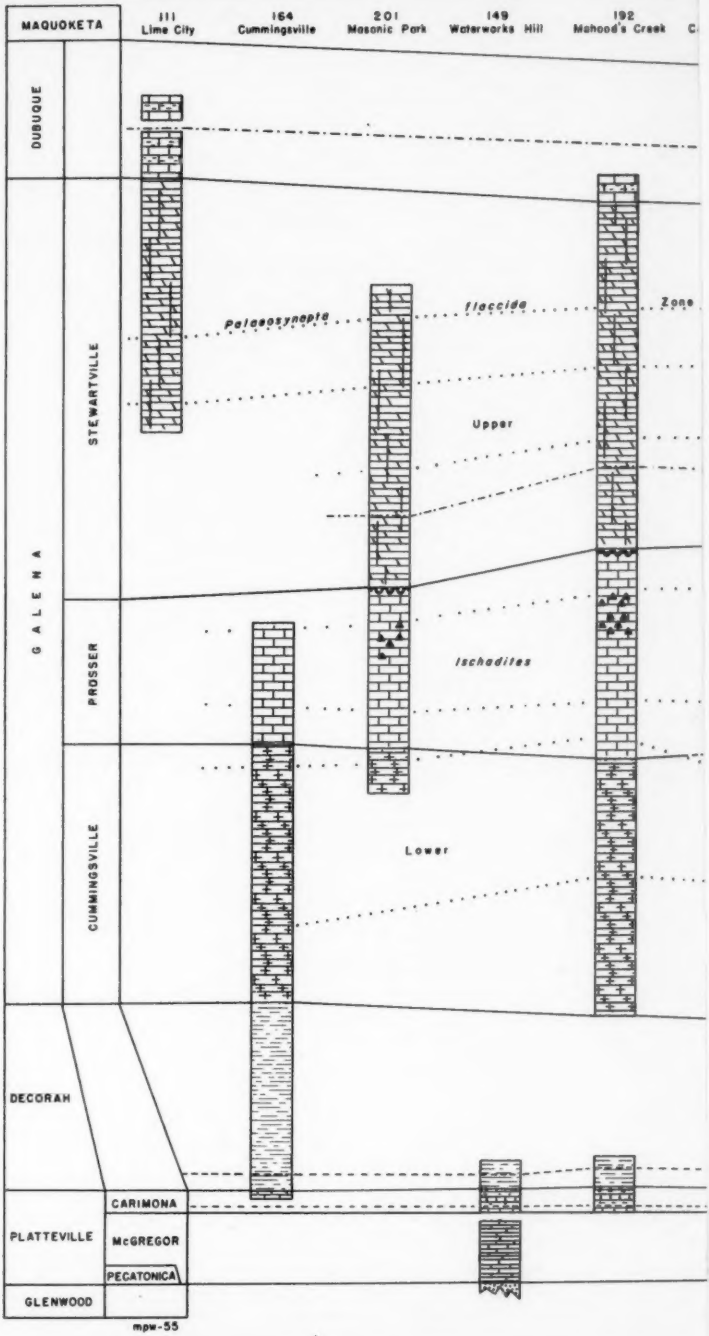

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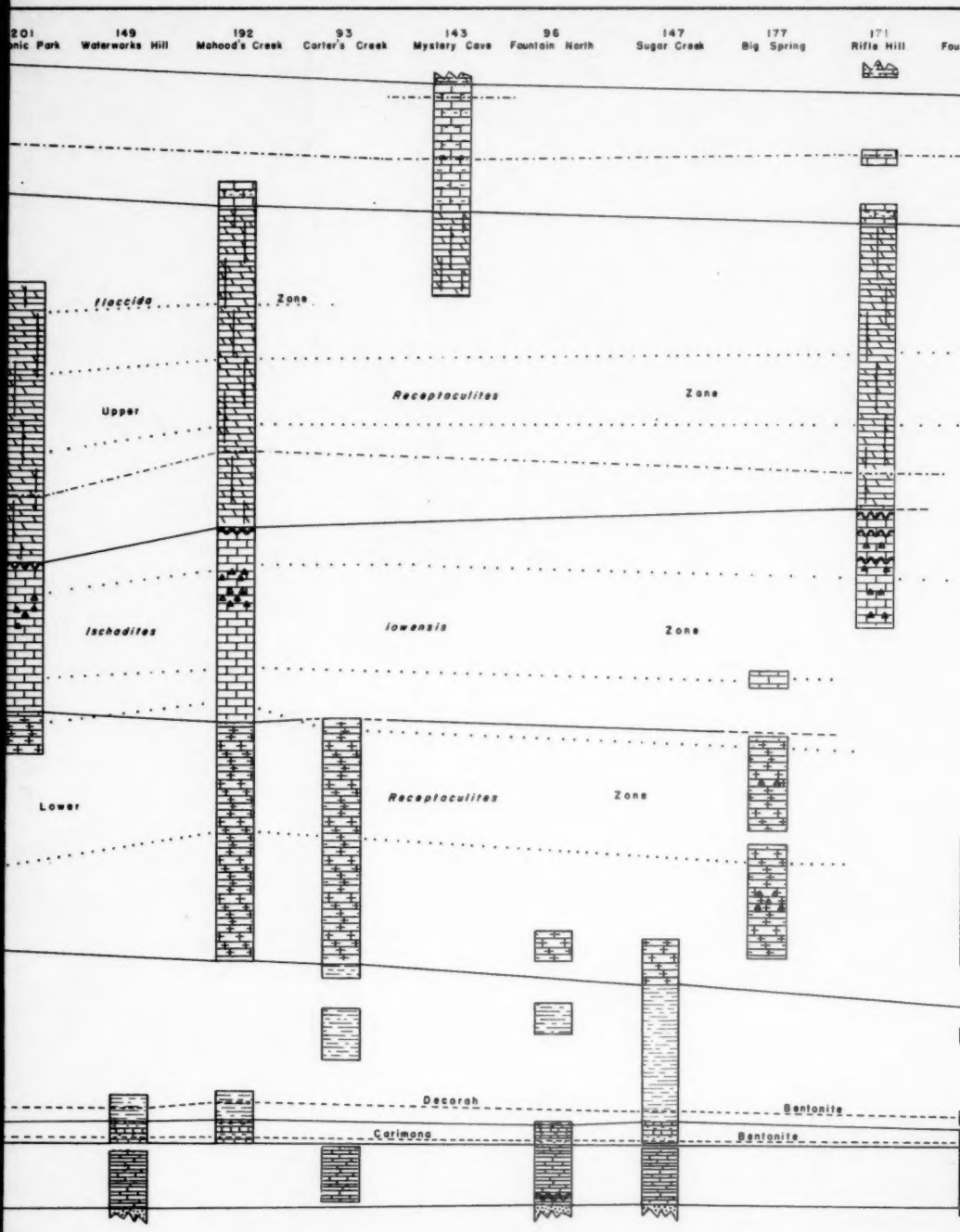

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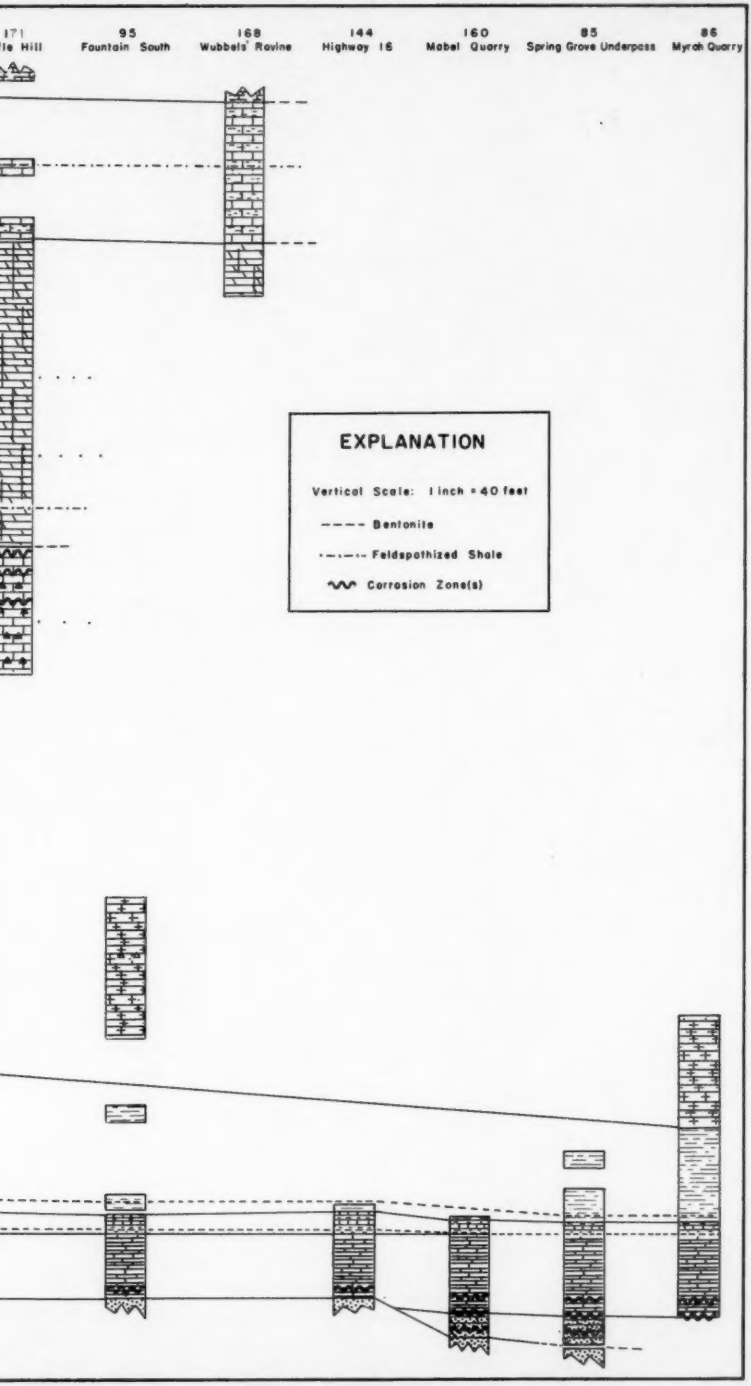

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Short Notes

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STRUCTURE OF LOPOLITHS: DISCUSSION

BY EUGENE N. CAMERON AND HARRY E. ABENDROTH

H. D. B. Wilson's interesting article on *Structure of lopoliths* (1956) prompts the writers to offer a few comments based on field observations of the Bushveld Complex during the past 5 years.

Wilson discusses variations in the thickness of the ultrabasic rocks (Critical and Transition Zones of Hall) along the strike of the "norite" portion of the lopolith. He relates the variations to alternating bulges and depressions in the floor of the complex; the ultrabasic rocks are thickest in the depressions and thin or even pinch out over the bulges. He concludes that in the deeper portions of the basin along the east-west axis the layered ultrabasic section is well developed and approximately as thick as the norite of the Main Zone, but that farther out on the flanks of the basin the intrusion steps upward in the sedimentary series, the deeper ultrabasic layers are gradually cut out, and even the Main Zone norite thins. He infers that the intrusion is not sill-like, the bottom dips more steeply than the layering, and the ultrabasic rocks are concentrated at the bottom of the intrusion.

During 1951–1955 the writers, assisted by D. L. Welch and B. L. Berman, mapped in detail (scales of 1:10,000 or larger) Farms Scheiding, Jagdlust, Winterveld 343, Mecklenburg, Hendriksplaats (with R. A. Bell), Annex Grootboom, and part of Grootboom, examined parts of Farms Maandagshoek, Mooihoek, Onverwacht, Doornbosch, Winterveld 424, Goudmyn, and Spitzkop with some care, and made a reconnaissance south from Annex Grootboom as far as Dwars Rivier and De Grooteboom. The locations of these farms, all in the Eastern Transvaal, are shown in Figure 1. The results of this work have some bearing on Wilson's conclusions.

Between Scheiding and Steelpoort, the thickness of rocks below the Merensky Reef varies along strike. On Jagdlust the thickness is about 10,000 feet. On Mecklenburg the thickness can be calculated only roughly but is on the order of 9000 feet. Southward from Mecklenburg, the thickness calculated depends on what one includes in the Bushveld Complex. On the Olifants River sheet (1914), Hall showed the Bushvelt "norite" (including the ultrabasic rocks) as almost everywhere separated from the floor (Pretoria series) by a peripheral zone of "diabase" (Fig. 1). Oddly enough, on the general map of the Bushveld in Hall's memoir (1932), the basal contact of the Bushveld Complex with the Pretoria series has the geographic position of the "contact" between "diabase" and "norite" shown on the individual sheets. In discussing the complex (p. 308–309, 356) Hall makes it clear, however, that he regards the diabase as part of the overall intrusive and the base of the diabase as the true floor. He speaks, for example, of contacts between norite and sediments of the Pretoria series, *e.g.*, in the eastern part of Moeyelyk. He states that on Farm Mooihoek and elsewhere one can pass without a break from norite into a diabase sill. In mapping on Jagdlust, Winterveld 343, Mecklenburg, and Hendriksplaats, the writers have found no break between the "diabase" and the overlying ultrabasic rocks. The two together correspond to Hall's Transition and Chill zones. The predominant exposed rocks of the zone shown by Hall as "diabase" on Mecklenburg are pyroxenites and mafic norites.

The difference, with respect to Wilson's concept, is best seen on Hendriksplaats and adjacent Derde Gelid. If we follow Hall's general map and take the floor of the Complex at the norite-diabase contact shown on the

Sekukuniland sheet (Hall, 1911), the thickness of rocks below the Merensky Reef on Hendriksplaats is about 8000 feet, and the floor transgresses the structure of the underlying rocks north and south of this farm. The rocks below the Merensky Reef then appear to thicken northward and southward into downward bulges of the floor. If the floor is taken at the base of the "diabase", however, the thickness of rocks above the floor on Hendriksplaats and adjoining farms, including three long inclusions of the Pretoria sediments, is between 11,000 and 12,500 feet—more than the thickness on Jagdlust and Mecklenburg. Such depressions as occur in the floor of the complex between the Olifants and Steelpoort rivers are now not downward transgressions across the footwall of the complex but simply places where the quartzites of the Magaliesburg formation of the Pretoria series are downwarped. These downwarps die out westward below the Merensky Reef but affect rocks as high as 2500 feet below the Reef.

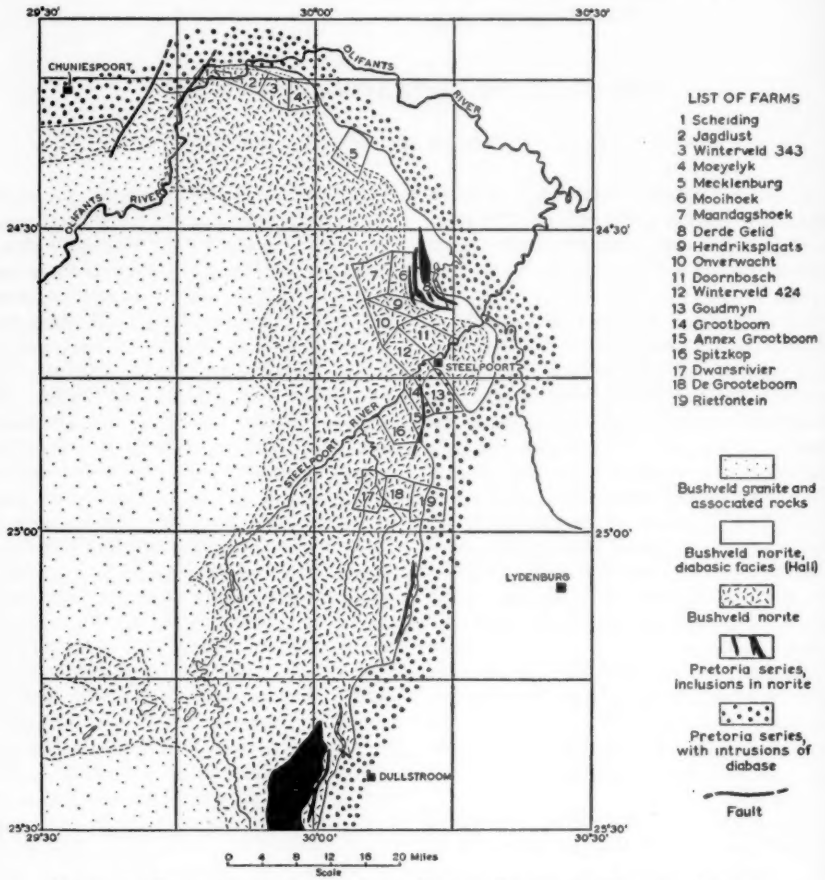

FIGURE 1.—GENERALIZED GEOLOGIC MAP OF EASTERN PART OF THE BUSHVELD COMPLEX
Adapted from Hall.

From the standpoint of gravitative accumulation of early-formed crystals, however, it seems likely that the *effective* base of the Bush-

veld Complex between Steelpoort and Mecklenburg would lie above all major inclusions in the lower part of the complex. In this sense one might speak of downward intrusive transgressions of the effective base of the complex. One would then expect that the sheets of norite between the long xenoliths of the Hendriksplaats-Derde Gelid area would approach the original average magma in composition and would differ from the ultrabasic series above the xenoliths. Hall gives evidence that this is so. The marginal parts of the peripheral "diabase" and its sill-like extensions beneath long xenoliths are described as noritic, gabbroic, and diabasic rocks, rather than as ultrabasic rocks.

It seems clear that large xenoliths or groups of xenoliths in the Transition Zone may inhibit the development of ultrabasic rocks. Relations in the Hendriksplaats area appear to offer one example, and there are others. Thus the classic display of rifted pyroxenites and peridotites on Farm Jagdlust (Malips Drift locality of Hall) breaks down southward against an assemblage of xenoliths on Farm Winterveld 343.

Wilson notes that south of Hendriksplaats the rocks below the Merensky Reef appear to thicken into the Steelpoort bulge. We doubt, however, whether any intrusive transgression is involved. The Lydenburg sheet (Hall, 1913) shows that the downwarp of the floor follows a syncline in a massive quartzite horizon of the Magaliesberg around the northeast and east sides of the bulge, but on the south side Hall shows the floor as transgressive to the sediments. Hall's mapping here and to the south needs revision. The quartzite forming the floor just south of Steelpoort is warped into a sharp anticline plunging to the north. The actual floor cannot be seen, but the pseudostratification of the basic rocks is conformable to this structure so far as can be determined from scattered outcrops. The apparent thickening of the ultrabasic rocks in the Steelpoort bulge seems to be due partly to the folding, partly to the fact that the ultrabasic rocks in general have lower dips (12° or less) than to the north.

South from the Steelpoort bulge the exposed thickness of ultrabasic rocks abruptly diminishes, but for a reason different from that in the north. Detailed mapping on Annex Grootboom and Grootboom and study of the rocks exposed on Goudmyn show that the ultrabasic rocks are dropped down along a steeply dipping major normal fault against the rocks of the floor. The displacement is so great that at one point an erosional outlier of the Merensky Reef abuts against the fault, indicating a displacement of the order of 8000 to 10,000 feet. The fault is part of a zone of disturbance (Hall, 1932, p. 210) that extends from Steelpoort southward at least as far as De Grooteboom, but the fault has not been traced to the south beyond Spitzkop.

For the area southward from De Grooteboom to Dullstroom, no detailed maps have been published. Hall's mapping (Lydenburg) suggests that ridges formed on the lower units of the ultrabasic series are partly cut off by a westward swing of the contact, at a point about 13 miles north of Dullstroom. Possible complications of the structure were also noted by Lombaard (1934). Similar swings of the contact east of Chuniespoort are due to transverse faulting, and the writers have been led to speculate whether a combination of transverse faulting and strike faulting, of which there is abundant evidence to the north, might be the cause of the disappearance of the ultrabasic series at this point, instead of the shallowing of the basin suggested by Wilson. Hall's maps suggest still another possibility—namely, lateral passage of the lower part of the norite into a series of sills, but Wilson (Personal communication) finds evidence against this interpretation.

The writers' data suggest one further comment. Wilson has described a structural contrast between the intricately folded and faulted sediments of the roof rocks and the gently warped Pretoria sediments of the floor and has made the interesting suggestion that the roof sediments have undergone a severe mountain-building epoch which has not affected the earlier sediments below. The writers' observations do not entirely support this. It is true that over long stretches the structure of the floor rocks is simple, but in some places these rocks are highly deformed. In the eastern Transvaal, for example, the sediments of the floor are highly folded in three areas—one extending south from the Steelpoort bulge, a second east of Mooihoek, and a third extending from Scheiding past Chuniespoort. Inasmuch as the Bushveld norite is itself deformed both in the first and third areas, there would appear

to be some question as to when the principal deformation of the floor rocks took place.

In this connection, the writers have noted that many inclusions in the lower part of the "norite" are intensely deformed. The structural competence of the rocks associated with the Bushveld Complex appears to be a factor. The simplicity of the floor in certain areas may be related to the occurrence of thick, massive quartzites in the Magaliesberg formation.

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STRUCTURE OF LOPOLITHS: A REPLY

By H. D. B. Wilson

The comments of Cameron and Abendroth on the writer's paper *Structure of lopoliths* are most interesting and add considerably to the published geology of the Bushveld Complex in the Steelpoort area. The writer wishes, however, to add a few further comments to help clarify some of the points of discussion.

Cameron mentioned that in a generalized paper it is difficult to determine how much field work had been done by the author before arriving at his conclusions (personal communication). The writer spent approximately 18 months in the field working on the Bushveld Complex in company with two other geologists. The field work was aided by a 1 inch equals a quarter of a mile aerial-photograph mosaic of the Bushveld intrusion and large portions of the wall rocks. Detailed airborne magnetometer maps of the intrusion were also available and were of great help in working out the structure because, for example, they not only show where transverse faults do occur but also show where they do not occur where magnetic horizons are not interrupted.

Obviously an intrusion as large as the Bushveld Complex cannot be completely mapped in 18 months, and considerable reliance must be placed on published maps and reports. Because the survey was of an economic character, certain selected portions of the intrusion were mapped and examined in considerable detail. Particular attention was paid to the lower contact of the lopolith, the lower layers of the intrusion, the Merensky horizon, the sulfide pipes in the western Bushveld, and to possible projections of the intrusion into the floor rocks. Most of the work was concentrated in the western Bushveld, but the area between Wanhoop and Tanteldoos in the eastern Bushveld near Dullstroom was carefully examined because of the apparent disappearance of the chromite and Merensky horizons and because Hall (1932) shows a large projection of the intrusion into the floor rocks. This area lies south of the area where Cameron and Abendroth and associates did their detailed mapping. Their area was examined only in a reconnaissance manner by the writer, but he was aided by the air photographs and the magnetometer maps, and by visiting Doctor Willemse of the Geological Survey of South Africa who was mapping in the Steelpoort bulge area.

The location of the floor of the Bushveld Complex is a difficult problem in many areas because of the scarcity of outcrops and the complication introduced by the similarity of the diabase zone of the main intrusion, and the older diabase sills in the Pretoria Series. Cameron and Abendroth point out that, north of Steelpoort, Hall has shown the floor in two positions on different maps. Cameron and Abendroth use the position on Hall's Sekuku niland Sheet (1911) which places the floor below several layers of "diabase" and Pretoria Series sedimentary bands. They point out that Hall (1932, p. 308–309, p. 356) regarded the diabase as part of the overall intrusion and the base of the diabase as the true floor. However, on one of these same pages, Hall (1932, p. 308) also states that the diabase zone of the main intrusion is "almost indistinguishable from those that build the sills within the girdle of the floor sediments." The writer therefore concluded that Hall's later map (1932) showed his final conclusion regarding the position of the true floor on farms Mooihoek, Derde Gelid, and Hendriksplaat. As Cameron and Abendroth point out, if we use Hall's later map, the floor transgresses the structure of the underlying rocks, but if the floor is taken as the base of the "diabase" and the sedimentary formations taken as xenoliths, there may not be transgression.

The writer used Hall's (1932, p. 210) conclusion that the Steelpoort bulge was formed by the basic magma breaking into the floor and transgressing the floor rocks because this was the only published description. Hall's conclusion was checked by examination of one area in the base of the bulge where ultrabasic rocks containing disseminated copper and nickel sulfide minerals lie in contact with Pretoria Series

sediments. At this location the normal unfaulted contact between the ultrabasic rocks of the Bushveld Complex and the floor rocks of the Pretoria Series dips somewhat steeper than the average layering of the eastern Bushveld Complex. Here, it did not appear that the bulge was caused by a flattened dip of the contact or layering, and also at this point as at many other places the diabase chill zone does not occur.

Hall's conclusion with respect to the structure of the Steelpoort bulge appeared reasonable in view of his careful description of the transgressive relations further south and with the detailed examination made of the floor of the intrusion near Tanteldoos in the Dullstroom sector. Hall's (1932, p. 209–210) description of intrusive transgression 10 miles south of the Steelpoort bulge has a distinct bearing on the problem:

> "A similar case of intrusive transgression is found 16 miles west-north-west of Lydenburg; here the highest of the four main Magaliesburg quartzites making the backbone of the Steenkampsbergen, after persisting many miles from Belfast northwards, comes to an abrupt end in a prominent peak that lies round the common beacon of the farms Vygehoek, Modderspruit, and Naauwpoort. The continuity on the east of this hill of the next lower quartzite does not bear out any possible faulting, while the distribution of the surrounding formations and their locally disturbed position point to the intrusion of the basic magma across part of its floor."

Hall, therefore, considered the possibility of faulting, as suggested by Cameron and Abendroth, but found evidence to the contrary.

Hall's description of this transgressive relation is similar to that in the area mapped by the writer and associates between Wanhoop and Tanteldoos about 15–25 miles farther south along the floor of the intrusion (northeast of Dullstroom on Cameron and Abendroth's Fig. 1). In this area the chromite and Merensky horizons disappear from the norite phase and are not found farther south. Hall (1932) also shows a large projection of the main intrusion into the floor. On the map this projection appears somewhat similar to some of the "offsets" at Sudbury, Canada. An examination of the air photographs showed that this "norite" projection is composed of bands of sedimentary rocks and basic sills. The photographs and ground mapping showed that these sediments can be traced back into the floor rocks where they form part of the normal Pretoria Series, and the basic intrusions are diabase sills intruding the Pretoria Series. Mapping across the mouth of Hall's bay of "norite" showed that it is not a norite bay but rather a part of the Pretoria Series cut by many of the older diabase sills. The main zone of the Bushveld intrusion is chilled against the sedimentary rocks and the older diabase on many outcrops, and the sediments and diabase show thermal metamorphic textures. The chilled contact shows the Bushveld intrusion is transgressing the older sediments and diabase and also shows that the transgression is not due to faulting. Hall's projection of the main intrusion into the floor in this area is another example of the uncertain position of the floor of the intrusion because of confusion with the older "diabase" sills in the Pretoria Series.

The location of transverse faulting is shown especially well by the air magnetic maps where the faults cross the magnetite horizons and some of the magnetic horizons of the Pretoria Series. At the Chuniespoort fault locality mentioned by Cameron and Abendroth the air magnetic map and the photographic mosaic show clearly that three faults are involved in causing the displacements of the Bushveld Complex horizons and contacts. These swings of the contact here are certainly not intrusive transgressions.

The magnetic maps show displacements of the magnetite horizons near the Steelpoort river between lat. 24° 52′ and 25° 5′ S. Probably the north easterly flowing portion of the Steelpoort River follows a zone of transverse faulting as shown by the blocky displacements of the magnetite horizon, the topographic lineament of the Steelpoort River, and the faulting described by Cameron and Abendroth in the Steelpoort bulge.

The magnetic maps show just as clearly, however, that transverse faults do not intersect the magnetite horizons south of lat. 25° 5′ S. because the magnetite horizons continue uninterrupted to latitude 25° 45′ S.—that is, about 16 miles south of the south boundary of Cameron and Abendroth's Figure 1. Thus in the southern sector of the eastern Bushveld the magnetite horizons are continuous on the west side of the floor of the Bushveld Complex, and

the formation described by Hall and quoted above is continuous on the east side of the Complex, so that transverse faulting cannot be invoked to explain the westward swing of the floor and the disappearance of the chromite and Merensky horizons near Dullstroom as suggested by Cameron and Abendroth. This lack of faulting, the chilled contact exposures, the abrupt thinning of the intrusion and thickening of the sedimentary series, with members truncated to the north, indicate intrusive transgression similar to that described by Hall a few miles farther north at the locality west-northwest of Lydenburg. Cameron and Abendroth have shown that the Steelpoort bulge is more complicated than described by Hall, but the relations seem clear farther south.

The writer should have made it clearer, in regard to the difference in structure between the underlying Pretoria Series basin and the overlying roof rocks, that, although a few sharp contortions occur in the Pretoria Series basin below the intrusion, these are comparatively small local wrinkles on the remarkably uniform, gently dipping, basin structure of the floor rocks; north of Beestekraal the Pretoria Series rocks overlying the intrusion are highly contorted, even isoclinally folded, and complexly faulted, and are somewhat comparable in structure to many of the highly contorted Canadian early Precambrian areas which have undergone intense mountain building.

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FUNNEL-SHAPED LAYERED INTRUSIONS

BY L. R. WAGER AND G. M. BROWN

In a recent paper Wilson (1956) has drawn attention to certain features of five funnel-shaped layered intrusions which he has examined in the field: the Bushveld, Great Dyke, Insizwa, Kungwe Bay, and Sudbury intrusions.

The intrusions considered by Wilson are called lopoliths but yet are described as being shaped like a funnel. The writers do not agree with this use of the term lopolith, which was defined by Grout (1918a, p. 518) as "a large lenticular, centrally sunken, generally concordant, intrusive mass, with its thickness approximately one-tenth to one-twentieth of its width or diameter". If the intrusions are shaped like a funnel or inverted cone, then they resemble the Peekskill norite for which Balk used the term funnel or "Trichter" pluton (Balk, 1927). When dealing with the Skaergaard layered intrusion, which could clearly be seen to have the shape of an inverted cone, Wager and Deer (1939, p. 52–59) adopted Balk's term funnel intrusion. Because of the analogy between the mechanism of cone-sheet intrusion (Anderson, 1936; Bailey *et al.*, 1924, p. 11) and that of the Skaergaard cone-shaped pluton Wager and Deer considered using the term cone pluton, but this was given up to avoid confusion with the term cone-sheet intrusion. Salomon (1903, p. 310) used the term "ethmolith" (literally, funnel rock) to describe certain tonalite intrusions in the Adamello massif of the Alps, but this far-from-euphonious word has not come into general use. The formation of funnel-shaped intrusions and cone sheets involves the upward removal of country rock to provide space, and both are presumably a response to high magmatic pressures in contrast to the lower pressures associated with piecemeal stoping or the sinking of larger blocks bounded by outward-dipping ring-shaped fractures.

If the five large basic layered intrusions which Wilson has described have the shape of inverted cones then the writers suggest that they be called funnel intrusions or ethmoliths. On the other hand if further study shows that some have been emplaced by a local downsag of the crust as illustrated by Grout (1918a, Fig. 2), then the term lopolith would remain appropriate. In view of the significant relationship between the form of the intrusion and the mode of emplacement it is important that the two terms lopolith and funnel intrusion should be kept available and used, when there is sufficient evidence, in accordance with the original definitions, for one implies a downward displacement of crustal material, the other an upward displacement.

When Grout introduced the term lopolith, he assumed that the floor of the Duluth gabbro was parallel to the banding or layering within the gabbro mass (1918a, p. 521). Recently Taylor has re-examined the Duluth gabbro near the city of Duluth and found that at Bardon's Peak (1956, p. 51) the lower contact of the layered series troctolite with Keweenawan basalt flows dips 45°E., whereas the banded structures within the troctolite dip at only 20°–30°E. Perhaps this may be taken as a general feature of the Duluth layered intrusion, and if so the Duluth gabbro itself will not be a lopolith but a funnel-shaped intrusion as envisaged by Wilson for the five cases he has considered. It may therefore be that the intrusion that led to Grout's conception of a lopolith will prove to be funnel-shaped, and if so it will not be possible to use it as the type example.

For layered intrusions the writers agree with Wilson that it is not safe to assume that the layering is parallel to the floor of the intrusion; indeed, observations seem to show fairly clearly that in some cases the dip of the layering is not parallel to the floor. Undisturbed layered intrusions seem to be composed of rocks arranged in layers which dip toward the center at a low angle, decreasing from about 20° near the margins to horizontal near the center. The layers may be compared with a stack of saucers, of steadily increasing diameter upward. The floors of the intrusions, on the other hand, are funnel-shaped, and so the central quasi-horizontal layers do not conform to the floor.

The relationship between the walls and the banded structures of the Skaergaard intrusion may be studied in detail, because of good exposure in a region of high relief. For about 30 meters from the margin the rocks of the border

The layered series of the Skaergaard intrusion was formed by crystals settling from a convecting magma. This has resulted in the deposition of essentially horizontal layers of crystal sediment at the bottom of the liquid;

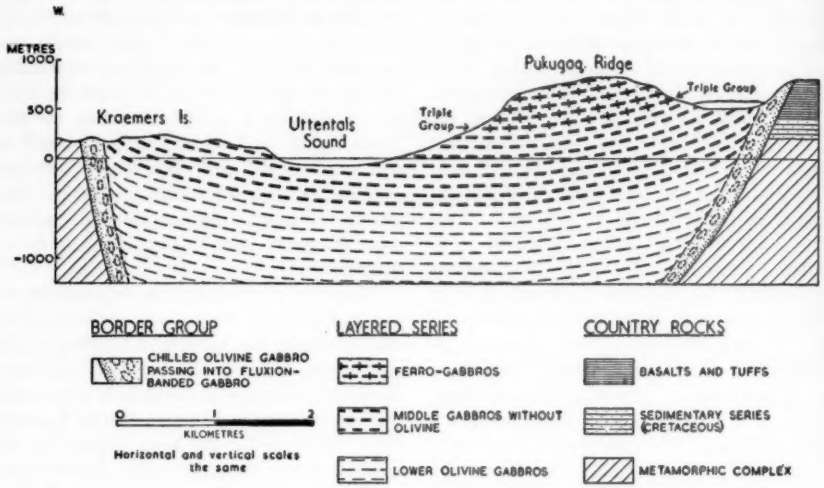

FIGURE 1.—GENERALIZED EAST-WEST SECTION OF THE SKAERGAARD INTRUSION, GREENLAND

group are fine-grained because of chilling; the main part of the group consists of variable, coarse gabbros with fluxion banding (Wager and Deer, 1939, p. 51–63). The fluxion structures of the marginal border group are parallel to the contact, and their character indicates upward or downward flow. Although now tilted, the layering is believed to have been almost horizontal at the time of formation, except that dips increase to about 30° as the border group is reached. The layering was thus originally saucer-shaped, and there is good igneous lamination due to the arrangement of the two longer axes of the tabular crystals of the rock parallel to the plane of layering. The igneous lamination is distinct from the fluxion structure of the border group, both in character and direction of dip. These structural relations were fully confirmed during a further visit by the writers in 1953, and they are summarized in the generalized east-west section (Fig. 1). The north-south section of Wager and Deer (1939, Pl. 26) is complicated because of a regional flexure of the crust which gives a southerly tilt to the whole mass.

the form of the layers was not influenced by the shape of the chamber walls except for some banking up of the crystal accumulate against the border group, giving the characteristic saucer shape to the layers. The border group, on the other hand, is considered to have formed from the convecting magma as a result of heat loss through the walls. The convecting magma was moving parallel to the walls, and the narrow zone of border rocks is characterized by structures whose directions are controlled by that of the walls. In a funnel-shaped intrusion of the Skaergaard type it is, therefore, necessary to identify two types of internal structure, the fluxion structures of a border group parallel to the walls of the intrusion, and the banding and igneous lamination of the layered series conforming only by chance to the direction of the walls.

If Wilson is right about the funnel-shaped form and the saucer-shaped layers of the five large basic intrusions he has studied, then they conform in these respects to the Skaergaard intrusion and not to the lopoliths as Grout conceived them. However, the intrusions de-

scribed by Wilson apparently differ from the Skaergaard intrusion in having no unit homologous to the border group. It may be that there are two types of funnel-shaped layered intrusions: one in which there is a border group, and another in which this group is absent and the crystal accumulate is banked against the chilled margin. In the Skaergaard magma chamber there was a single system of convection currents during most of the differentiation period, and the existence of the border group is believed to be due to this. In other intrusions there may have been no convection currents but direct sinking of crystals so that the crystal precipitate is banked against the outer walls.

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TERMINOLOGY OF LAYERED INTRUSIONS

BY H. D. B. WILSON

In the recent paper (Wilson, 1956) referred to by Wager and Brown the writer avoided any discussion of the terminology of the intrusion types because he wished to lay all stress on the structural type. He used the term lopolith because Grout (1918, p. 521) classified the Sudbury intrusion and the Bushveld complex as lopoliths, the term he was defining.

Grout (1918) did not imply any genetic significance to the term lopolith; he states clearly that he is referring only to form. He says (p. 516) "several other larger intrusions are of similar form, and it is here suggested that the form deserves a special name;" and again (p. 518) "such a name is better based on known facts of form or relations than on any theory of origin, and the name proposed is lopolith." It seems, therefore, that Grout did not imply the genetic difference between lopoliths and funnel intrusions as suggested by Wager and Brown.

In general, the writer agrees with Brown and Wager that funnel intrusion is the best name for the intrusions described by the writer. The difficulty seems to be that, where such an intrusion enters a flat-lying series of rocks, the funnel shape may become so flattened that it may be recognized only by detailed mapping, and the floor of the intrusion may be conformable for long distances as is the case with the Bushveld complex.

Generalizations cannot be made about the border or chill group of most of the large intrusions discussed by the writer because of the scarcity of outcrops of the contacts and border zones. The chill zone of the western Bushveld complex is very rarely exposed and is largely mapped by the presence of red soils. The rare outcrops that do occur are not close to the contact with the floor rocks. The contact of the Great Dyke with its walls is exposed in only three small outcrops, and here the intrusion is serpentinized. The contact area is generally beneath an alluvium- or soil-filled valley. The only good exposures near the walls are in some of the chrome workings, but these workings never reach the walls. At Insizwa Mountain the floor of the intrusion is covered by a rubble zone, and, while it usually can be located within 100 feet, the actual contact is not exposed. The contact was seen only in Waterfall gorge where a chill zone about 15 feet thick has been exposed in the mine workings. This chill zone follows the dip of the contact rather than the layering. The zone was not examined for fluxion structure in the mine.

The contact of the Sudbury intrusion is well exposed. Chill phases of the main norite generally are not present except along the north range where it has been observed in many drill holes. Platy flow structures caused by clots of labradorite have been observed in many places, particularly along the east range. These clots dip inward toward the center of the intrusion parallel to the dip of the main norite contact which along the east range dips fairly uniformly inward at about 60° as shown by many drill holes. The contact of the Sudbury intrusion is complicated in many places by a fine-grained phase of the intrusion which occurs in irregular bulgy bodies around the edge of the intrusion. This phase which is locally termed "quartz diorite" is commonly noritic and contains numerous xenoliths and sulfide blebs and masses. The "quartz diorite" intrudes the main norite phase so that it cannot be a chill zone.

Fluxion structures are common in the central part of the main Sudbury norite. They are well exposed in the Creighton sector where linear fluxion structures dip northward toward the center of the intrusion at an angle of about 40°—that is, approximately parallel to the dip of the frozen bottom of the intrusion along the south range.

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